WORKSHOP ON

TECTONIC EVOLUTION OF GREENSTONE BELTS

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WORKSHOP ON

TECTONIC EVOLUTION OF GREENSTONE BELTS

Edited by

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and

Lewis D. Ashwal
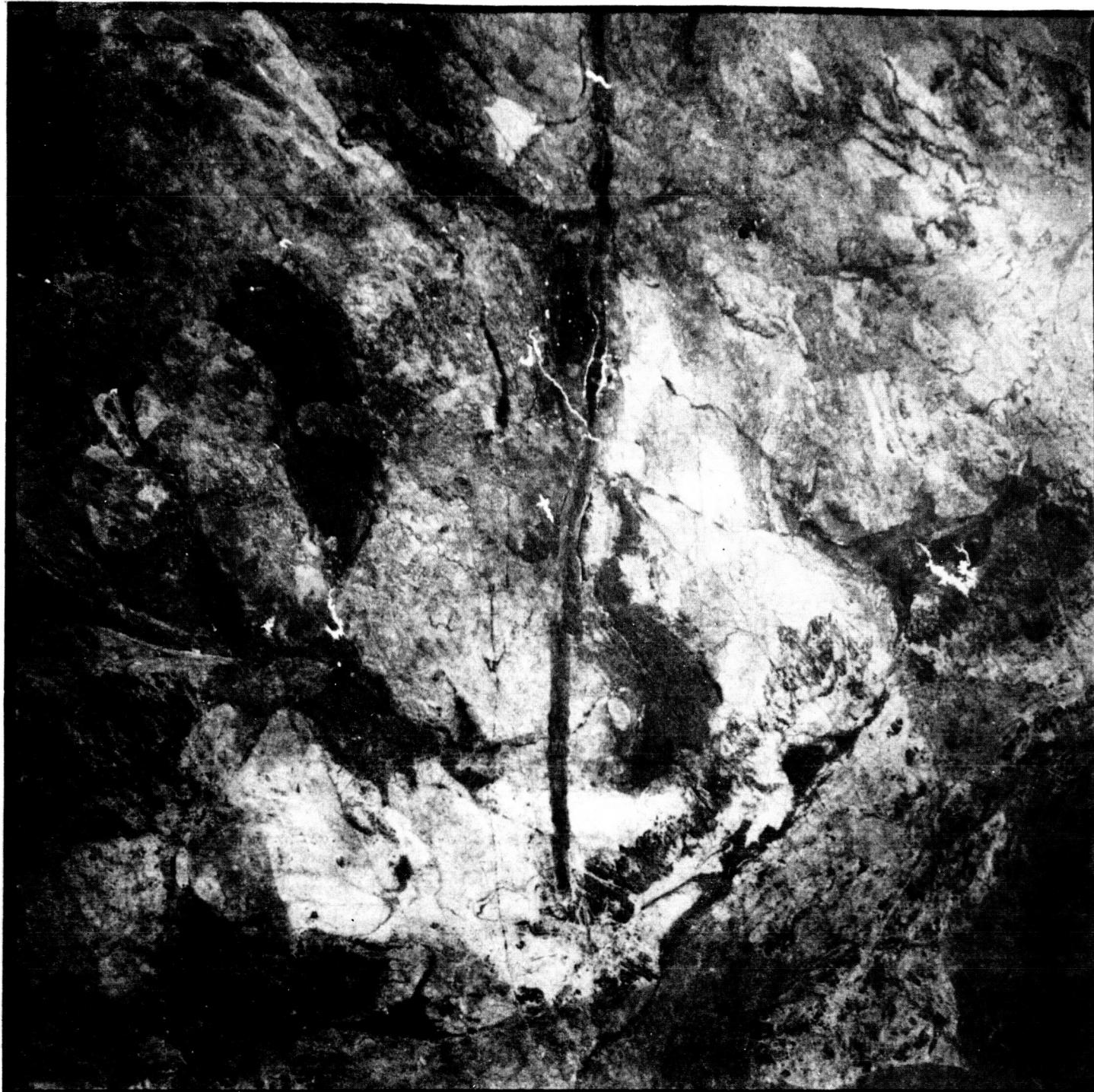
LEGEND

GREENSTONE BELTS

BELINGWE	Ax, Ay
BUHWA	By
BULAWAYO	Cx
FILABUSI	Dx, Dy
GWANDA	Ex, Ey
GWERU	Fx, , Fz (Fy too narrow to show)
MANESI	Gx
MASUINGO	Hx, , Hz
SHANGANI	Jx, Jy
SHURUGWI	, , Kz
OTHERS	L

x : Upper Bulawayan, circa 2.6 - 2.7 Ga.
y : Lower Bulawayan, circa 2.7 - 2.9 Ga.
z : Sebakwian , $>$ 3.45 Ga.*

* A recent zircon from the Wanderer Conglomerate at Selukwe has
  yielded an approximate age of 3.8 Ga.

MASHABA ULTRAMAFIC SUITE

undated, presumed upper Bulawayan (~ 2.6 - 2.7)

MASHABA COMPLEX	M1
SHABANI COMPLEX	M2
OTHERS	M3
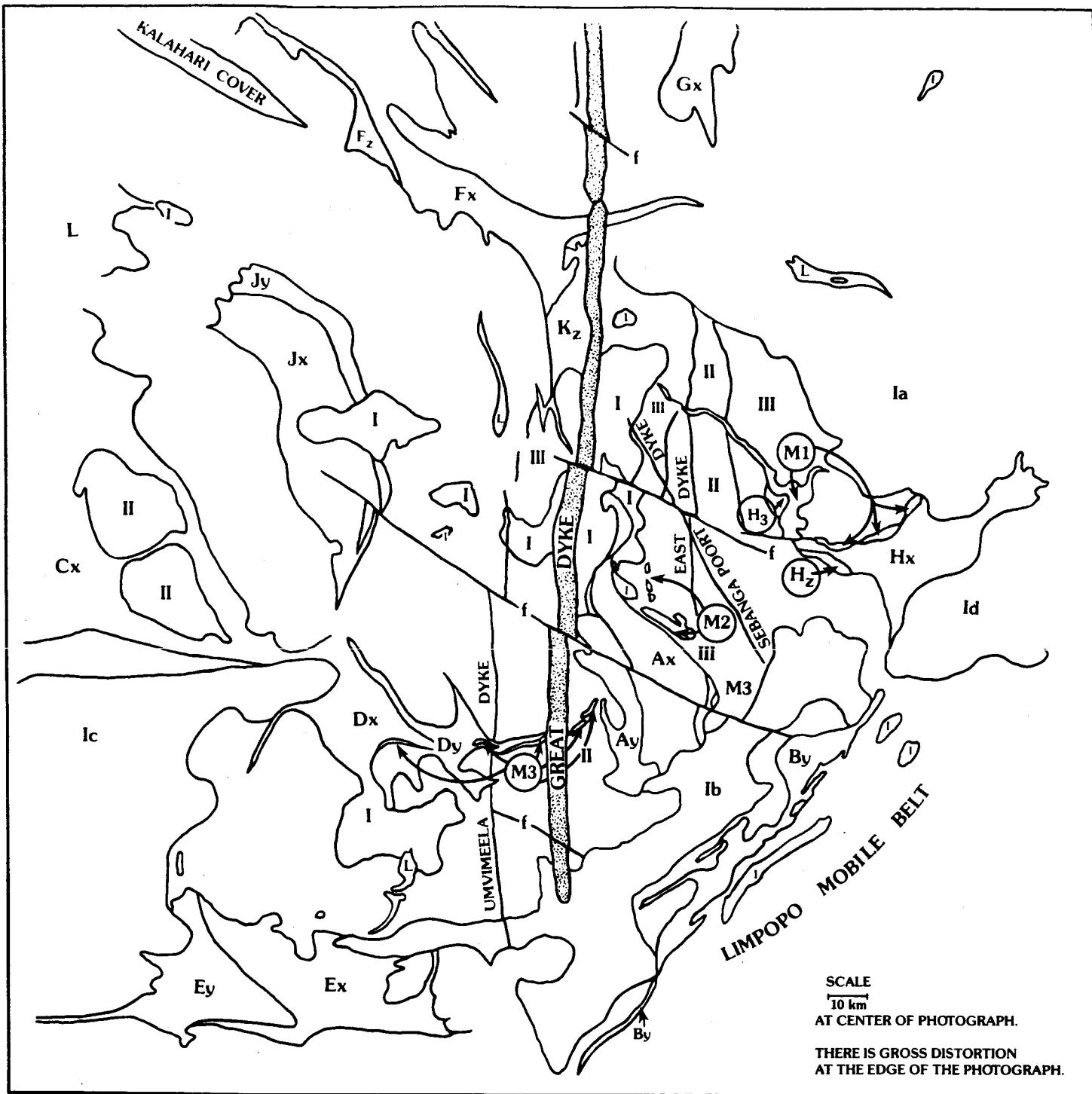
GRANITOIDS   I

CHILIMANZI SUITE

CHILIMANZI   Ia
CHIBI   Ib
MATOPOS   Ic
ZIMBABWE   Id

composite Rb/Sr isochron from various localities
around the craton: 2570 ± 25 Ma.

TONALITES   II

ANCIENT GNEISSES   III

undifferentiated, circa 3.5 Ga.
(Shibani shaft : 3490 ±120 Ma Rb/Sr).

Contents

Introduction

Background

Greenstone belts include some of the oldest rocks on Earth; consequently, they can provide us with direct evidence of Earth processes taking place as much as 3.7 Ga ago, and represent a foundation with which to begin an understanding of prior events during the first 800 Ma of Earth history. Greenstone belts also provide an anchor for those who wish to extrapolate back in time to understand the secular variations and/or punctuated patterns of 3.7 Ga of Earth evolution represented in the rock record. Despite this important aspect of greenstone belts, they are relatively little understood: there is not even a good definition of greenstone belts, and there is no agreement about the time span they occupy during Earth history. Until recently, greenstone belts were thought of as extremely thick stratigraphic sections of mafic-ultramafic lavas, with lesser sediments and felsic volcanics, metamorphosed at greenschist facies, and engulfed as synclines in a sea of intruded granitoid plutons. Their overall shape and tectonic style was thought to be due to density inversions and diapirism, the inferred result of an unstable configuration of sialic crust overlain by a dense simatic layer. Vertical Archean tectonics reigned over a lenghthy period during which the term "schist belt," as previously applied to greenstone belts, faded into the background. This enhanced the impressions that greenstone belts had been little affected by the sort of penetrative strains and tectonism that we are accustomed to seeing in Phanerozoic mountain belts. Over the last decade, however, the simplistic models of greenstone belts have not stood up to an onslaught of modern structural and geochemical investigations in many parts of the world. The message that has slowly emerged is that if we want to understand the secrets that greenstone belts hold to unraveling our planet's evolutionary blueprint, we must work hard for it: foremost in the field, with modern follow-up work in well-equipped laboratories. Greenstone belts are now being shown to be extremely complicated, far more so than even the staunchest uniformitarians would have predicted ten years ago. Foremost in revealing their original stratigraphic and geochemical signatures are structural and tectonic studies—those that endeavor to unstrain and restore the rock components of these belts to their primary dispositions. Understanding the tectonic evolution of greenstone belts will open vast horizons for those interested in the Archean siderosphere, asthenosphere, lithosphere,

hydrosphere, atmosphere, and biosphere. The rationale of the workshop was to set these tectonic studies on firmer footing. The meeting was part of NASA's Early Crustal Genesis Project, administered by the Lunar and Planetary Institute.

Outline and Structure of the Workshop

It was decided at the outset that the focus of the workshop would be "tectonics," broadly defined, and contributors were encouraged to integrate their ideas and data accordingly. The conveners invited ten researchers to team up in four groups to present four keynote review talks and highlight significant problems and new avenues for solutions. Both poster presentations and student participation were strongly encouraged.

The workshop was attended by eighty-six scientists from six continents. Results of research on greenstone belts from most major Archean cratons were presented: North and South America, Africa, India, and Australia (Fig. 1). About half of the contributions were in the form of posters, and about 15% of the participants were students. Based on the abstracts submitted, the oral program was divided into three sessions: (1) rock components, sources, provenances, and structures; (2) magmas, heat flow, fluids, and strain; and (3) boundaries, surrounding rock terranes, and their relationships. Lengthy discussion times (30–45 minutes) were interspersed between invited and contributed talks. All of these sessions were recorded on videotape. Well-lubricated evening sessions were reserved for viewing the poster presentations and for meetings of working groups, during which participants were asked to evaluate critically the data and ideas presented, and outline future research plans. The final session consisted of working group reports, technical summaries by a few spokespeople selected by the conveners, and further extensive discussion.

Results of the Workshop

The conveners consider the following major points to have emerged from the workshop:

1. Most greenstone belts are severely tectonized.
2. Greenstone belts occur at metamorphic grades varying from subgreenschist to granulite facies.
3. A single tectonic environment applicable to greenstone belts does not exist; different tectonic enviroments can be found within

Pangea Circa 200 Ma.
Van der Grinten Projection
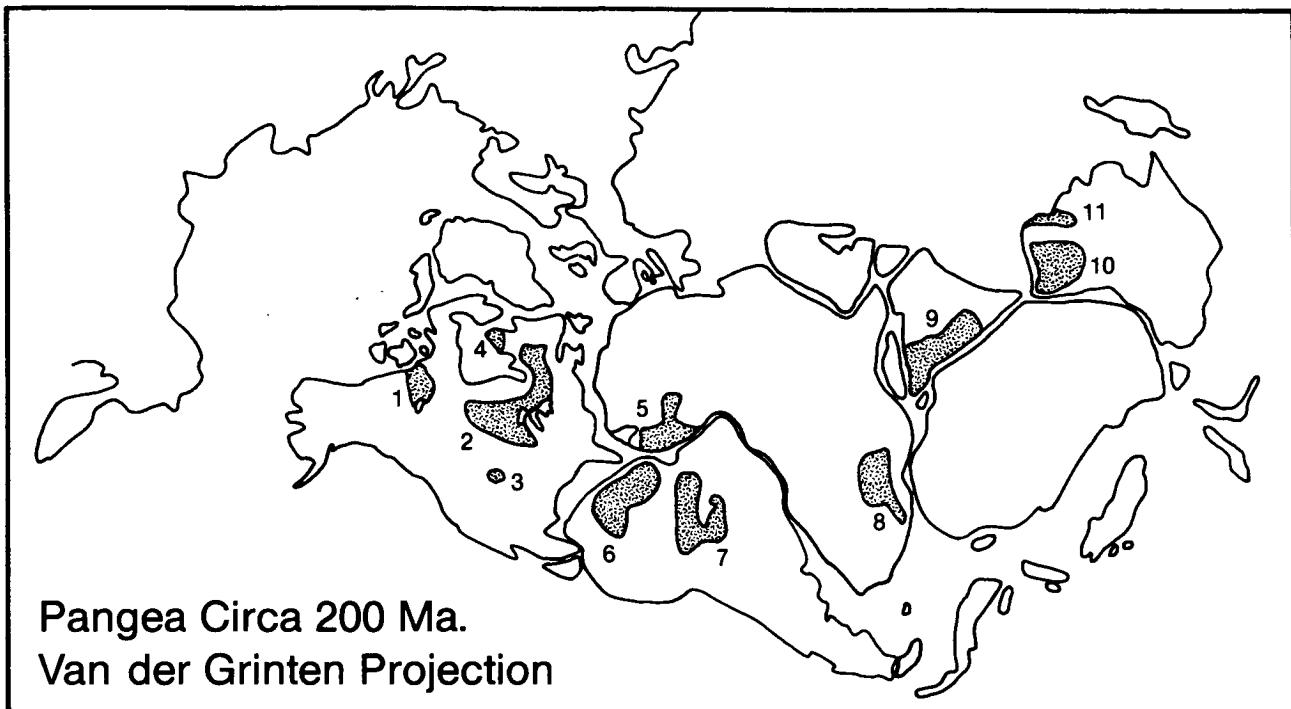
Fig. 1 Distribution of Precambrian terranes from which Greenstone Belts were described at this workshop.

1. Slave Province
2. Superior Province (Including Minnesota, USA)
3. Wind River Mountains (Wyoming)
4. Cape Smith Belt
5. West African Shield
6. Guyana Craton
7. Central Brazil Craton   } Amazonian Shield
8. Southern African Shield (Including the Zimbabwean Craton, the Kaapvaal Cratons, and the Limpopo Mobile Belt)
9. Indian Shield
10. Yilgarn Craton
11. Pilbara Craton

and among greenstone belts. Typically, greenstone belts contain mixtures of components from different tectonic environments.

4. Many of their features can be explained in the framework of plate tectonics.

5. Different cratons may be dominated by greenstone belts of broadly differing tectonic regimes.

6. The volume of komatiite in greenstone belts has probably been overestimated.

7. The MgO content of liquids parental to komatiites was likely less than 27%.

8. Compositions of rocks in many greenstone belts have been highly affected by metasomatism.

9. Overall structures of greenstone belts may not be related to granitoid diapirism; there may be a more subtle relationship between horizontal tectonics and granitoid plutonism.

10. Interpretations of critical field relationships are commonly equivocal.

Both presentations and discussions were highly stimulating and educational for all. An extensive summary of the technical sessions, prepared from the videotapes*, is included in this report. Also included are reports of the Working Groups, disciplinary summaries, and both invited and contributed abstracts. We hope this document will serve to foster new research on greenstone belts.

Maarten J. de Wit
Lewis D. Ashwal
Houston, Texas
March, 1986

*Videotapes can be viewed at LPI or copies purchased on request.

II. Program

SESSION I—Greenstone Belts: Rock Components, Sources, Provenances, and Structures
Chairman: L. D. Ashwal

Welcoming remarks
K. Burke*

Greenstone belt tectonics: Some relevant outstanding questions
M. J. de Wit* and L. D. Ashwal

Invited Keynote Talks

The rock components and structure of Archean greenstone belts
D. R. Lowe* and G. R. Byerly

Greenstone belts: Their components and structure
J. R. Vearncombe*, J. M. Barton, Jr., D. D. van Reenen, and G. N. Phillips

Discussion

Break

Contributed Papers

Preliminary structural model for the southwestern part of the Michipicoten greenstone belt, Ontario
G. E. McGill* and C. H. Shrady

Transpression as the main deformational event in an Archean greenstone belt, Northeastern Minnesota
P. J. Hudleston*, D. Schultz-Ela, R. L. Bauer, and D. Southwick

Archean wrench-fault tectonics in the Abitibi greenstone belt of Canada
C. Hubert* and J. N. Ludden

Discussion

Sedimentological and stratigraphic evolution of the Southern part of the Barberton greenstone belt: A case of changing provenance and stability
D. R. Lowe* and G. R. Byerly

Barberton greenstone belt volcanism: Succession, style and petrogenesis
G. R. Byerly* and D. R. Lowe

Discussion

SESSION I (continued)

2:00–5:30 p.m.

Chairman: W. S. F. Kidd

Dismembered Archean ophiolite in the southeastern Wind River Mountains, Wyoming: Remains of Archean oceanic crust
G. Harper*

Evidence for spreading in the Lower Kain group of the Yellowknife greenstone belt: Implications for Archean basin evolution in the Slave Province
H. Helmstaedt* and W. A. Padgham

Discussion

Zircon Lu-Hf systematics: Evidence for the episodic development of greenstone belts
E. Smith*, M. Tatsumoto, and R. M. Farquhar

Rhyolitic components of the Michipicoten greenstone belt, Ontario: Evidence for late Archean intracontinental rifts or convergent plate margins in the Canadian Shield?
P. J. Sylvester*, K. Attoh, and K. Schultz

Discussion

Break

The western Wabigoon subprovince, Superior Province, Canada: Late Archean greenstone succession in a rifted basement complex
G. R. Edwards* and D. W. Davis

A continental rift model for the La Grande greenstone belt
T. Skulski*, A. Hynes, M. Liu, D. Francis, B. Rivard, and K. Slamatelopoulou-Seymore

Discussion

Evidence for structural stacking and repetition in the greenstones of the Kalgoorlie District, Western Australia
J. E. Martyn*

A simple tectonic model for crustal accretion in the Slave Province: A 2.7-2.5 Ga "Granite-greenstone" terrane, N.W. Canada
P. F. Hoffman*

Discussion

Poster and Keg Session—7:00-11:00 p.m.

Friday, January 17, 1985

8:30 a.m.-12:30 p.m.
SESSION II—Greenstone Belt Externalities: Magmas, Heat Flow, Fluids and Strain
Chairman: K. Burke

Invited Keynote Talk
Greenstone belt tectonics—thermal constraints
M. Bickle* and E. G. Nisbet

Contributed Papers
Volcanologic constraints on Archean tectonics
P. C. Thurston* and L. D. Ayres

Komatiite genesis in the Archean mantle, with implications for the tectonics of Archean greenstone belts
D. Elthon*

Archean megacrystic plagioclase units and the tectonic setting of greenstones
   W. C. Phinney*, D. A. Morrison, and D. Maczuga

Discussion

The dehydration, rehydration and tectonic setting of greenstone belts in a portion of the Northern
Kaapvaal Craton, South Africa
   D. D. van Reenen, J. M. Barton, C. Roering, J. C. van Schalkuyk, C. A. Smith, J. H. de Beer, and
   E. H. Stettler*

Discussion

Break

Thermal implications of metamorphism in greenstone belts and the hot asthenosphere—thick lithosphere
paradox
   P. Morgan*

Hot spot abundance—ridge subduction, and the evolution of greenstone belts
   D. Abbott* and S. Hoffman

Discussion

2:00–5:00 p.m.
**SESSION III—Greenstone Belts: Their Boundaries, Their Surrounding Rock Terranes and Their
Interrelationships
Chairman: P. F. Hoffman**

Invited Keynote Talk
Greenstone belts: Their boundaries, surrounding rock terrains, and interrelationships
   J. A. Percival* and K. D. Card

Contributed Papers
New insights into typical Archean structures in greenstone terranes of Western Ontario
   W. M. Schwerdtner*

Correlations and contrasts in structural history and style between an Archean greenstone belt and adjacent
gneiss belt, NE Minnesota
   R. L. Bauer*, P. J. Hudleston, and D. L. Southwick

The stratigraphy of the Steep Rock Group, NW Ontario, with evidence of a major unconformity
   M. E. Wilks* and E. G. Nisbet

Discussion

Break

A continuous record of tectonic evolution from 3.5 Ga to 2.6 in Swaziland and Northern Natal
   D. R. Hunter, A. H. Wilson, J. A. Versfelt, A. R. Allen, R. G. Smith, D. W. W. Sleigh, P. B. Groenewald,
   G. M. Chutter, and V. A. Preston*

The Yilgarn Craton, Western Australia: A tectonic synthesis
   R. E. P. Fripp*

Geochemical characters and tectonic evolution of the Chitradurga schist belt: An Archean suture (?) of the Dharwar Craton, India
  *S. M. Naqvi**

Kolar schist belt: A possible Archean suture zone
  G. N. Hanson, E. J. Krogstad, V. Rajamani, and S. Balakrishnan*

Discussion

WORKING GROUP SESSIONS

7:00 p.m.–9:00 p.m.

Working Group I—Hess Room
Working Group II—Berkner Room

Saturday, January 18, 1986
9:30 a.m.–12:30 p.m.

SESSION IV—Research in Greenstone Belt Tectonics: Synthesis and Destiny
Chairman: L. D. Ashwal

Technical assessment by spokesperson(s)
  Working Group I

Technical assessment by spokesperson(s)
  Working Group II

Discussion

Summarizers

P. J. Wyllie: A Petrologic Viewpoint
D. W. Davis: A Geochronological Viewpoint
E. G. Nisbet: A Sedimentological Viewpoint
R. E. P. Fripp: A Structural Viewpoint
B. Gorman: An Ore Deposits Viewpoint
L. Losier: A Geophysical Viewpoint
P. Morgan: A Thermal Viewpoint
K. Burke: A Tectonic Viewpoint

*Denotes speaker.

Poster Presentations

7:00 p.m.–11:00 p.m.,
Thursday, January 16, 1986

(Open throughout the Workshop)

Group A—Northern Hemisphere—Reception Area, LPI

(1) Tectonic setting and evolution of Late Archean greenstone belts of Superior Province, Canada
  K. D. Card

(2) Geophysical characteristics and crustal structure of greenstone terranes, Canadian Shield
  M. D. Thomas, L. Losier, P. C. Thurston, V. K. Gupta, R. A. Gibb, and R. A. F. Grieve

(3) High precision U-Pb geochronology and implications for the tectonic evolution of the Superior Province
D. W. Davis, F. Corfu, and T. E. Krogh

(4) Age constraints on the evolution of the Quetico Belt, Superior Province
J. A. Percival and R. W. Sullivan

(5) Rainy Lake wrench zone: An example of an Archean subprovince boundary in Northwestern Ontario
K. H. Poulsen

(6) Deformational sequence of a portion of the Michipicoten greenstone belt, Chabanel Township, Ontario
C. H. Shrady and G. E. McGill

(7) A new 1:1,000,000 geologic map of Slave Province and early Proterozoic bounding origins
P. Hoffman

(8) Is the Cameron River greenstone belt allochthonous?
T. M. Kusky

(9) Basement-cover relations and internal structure of the Cape Smith klippe: A 1.9 Ga greenstone belt in Northern Quebec, Canada
M. R. St-Onge, P. Hoffman, S. B. Lucas, D. J. Scott, and N. J. Begin

Group B—Southern Hemisphere—Hess Room, LPI

(10) A continuous record of tectonic evolution from 3.5 Ga to 2.6 Ga in Swaziland and Northern Natal
D. R. Hunter, A. H. Wilson, J. A. Versfelt, A. R. Allen, R. G. Smith, D. W. W. Sleigh,
P. P. Groenewald, G. M. Chutter, and V. A. Preston

(11) Crustal structure of the Archean granite-greenstone terrane in the northern portion of Kaapvaal Craton
J. H. de Beer, E. H. Stettler, J. M. Barton, Jr., D. D. van Reenen, and R. R. Vearncombe

(12) Two contrasting metamorphosed ultramafic-mafic complexes from greenstone belts, the northern Kaapvaal Craton and their significance in Archean tectonics
C. A. Smit and J. R. Vearncombe

(13) The dehydration, rehydration and tectonic setting of greenstone belts in a portion of the Northern Kaapvaal Craton, South Africa
D. D. van Reenen, J. M. Barton, Jr., C. Roering, J. C. van Schalkwyk, C. A. Smit, J. H. de Beer,
and E. H. Stettler

(14) A mid-Archean ophiolite complex, Barberton Mountain Land
M. J. de Wit, R. Hart, and R. Hart

(15) Extensional tectonics during the igneous emplacement of the mafic-ultramafic rocks of the Barberton greenstone belt
M. J. de Wit

(16) Felsic igneous rocks within the Barberton greenstone belt: High crustal level equivalents of the surrounding tonalite-trondjemite terrain, emplaced during thrusting
M. J. de Wit and A. H. Wilson

(17) Polyphase thrust tectonics in the Barberton greenstone belt
I. Paris

Print Only Abstracts

III. Tectonic Evolution of Greenstone Belts: Some Relevant Outstanding Questions

Maarten J. de Wit and Lewis D. Ashwal

These are some relevant questions that emerge from the set of abstracts received for the workshop; these questions could be used to focus the discussions of the workshop. We would like the participants to assess these questions critically and to modify, reformulate, and if possible answer some of them during the workshop sessions. Can more important questions be formulated? These same questions could also serve as guidelines for the Working Groups when they meet to overview the data presented at the workshop and to outline avenues for future research in the tectonics of greenstone belts.

1. Should we be seeking a unified tectonic environment for all greenstone belts, or do greenstone belts represent a spectrum of tectonic environments? How can we identify those settings? Are greenstone belts part of more extensive orogens? Are greenstone belts allochthonous? How do we prove/disprove this?

2. Are the older Archean greenstone belts (i.e., Selukwe, Barberton, Pietersburg, Nondweni, Pilbara, Isua) different from the later Archean belts? If so, in what respect? There is very little compiled data on the quantitative estimates of rock types within the greenstone belts: a first-order objective. Moreover, because it is now clear that many greenstone belts are tectonic melanges, can we attach significance to the presence/absence of particular rock types in groups of greenstone belts?

3. Are greenstone belts thrust complexes; how important is late strike-slip motion? There is a growing data base indicating extensive deformation and tectonism in the form of faulting and complex folding. How do we recognize significant thrusting without fossils and resolvable chronological constraints? How reliable are the stratigraphic models that were predominantly built-up prior to the recognition of thrust-tectonics? How do we construct a reliable stratigraphy in the face of so few constraints?

4. Greenstone belts appear to have complex tectonic histories. Do they record Wilson cycles? If so, how would we recognize suture zones, particularly if greenstone belts are allochthonous? Are any greenstone belts fossil rifts (aulochogens)?

5. What do the sedimentary rocks in greenstone belts tell us about their tectonics? How can we determine the depth of deposition of Archean sediments? Are sedimentary rock assemblages in greenstone belts distinctive? What can their provenances tell us about earlier crustal history?

6. Do the simatic rocks of some greenstone belts contain ophiolites or fragments thereof? Sheeted intrusives have now been described from one early and two late Archean greenstone belts. Sheeted dykes are the most persuasive evidence for the presence of an ophiolite-like sequence in Phanerozoic orogens. Are gabbros or norites less common in these sections than in Phanerozoic ophiolites, and if so, does this indicate that intermediate stage magma chambers were less abundant during the formation of Archean oceanic crust? That might reflect more efficient spreading in the Archean relative to today.

7. Do the Archean simatic rocks represent Archean ocean crust or marginal basin crust? Can we use these rock sequences to reconstruct sections through the Archean oceanic crust, or do they merely represent obducted parts of specific oceanic tectonic regimes? Are there enough greenstone belts preserved to confidently reconstruct Archean oceanic-crust sections?

8. Komatiites: What is the tectonic significance of their abundance in the Archean compared to younger terranes? Are all komatiites extrusives? Could they also be sills or dikes? To what extent have they been affected by cumulus processes? What is the highest MgO content of liquids parental to komatiites (estimates range from 18–33%)? What was the degree of partial melting of the mantle to produce these komatiites (estimates range from 10–80%)? These problems have a great bearing on the temperatures of the Archean mantle and erupting of Archean volcanics and hence on Archean surface tectonics and lithosphere structures. Peripherally (for this workshop), it also bears on the problem of Archean asthenosphere physics: Was there a deep magma ocean?

9. Calcic, megacrystic anorthosites are another rock type unique to the Archean. Can these provide clues to the physics and chemistry of Archean magmatic processes? What was their tectonic setting?

10. Was Archean oceanic crust extensively hydrated during its formation or during later metamorphism? To what extent has the chemistry of this crust been altered by these processes? Are komatiite compositions tricking us? There are no completely fresh komatiites in greenstone belts—most are serpentinites. To what extent is their high MgO content a reflection of their igneous character rather than a metasomatic overprint? Are all spinifex textures igneous or are some metamorphic? How do we distinguish them?

11. What is the density of the simatic rocks in greenstone belts if the komatiites are in reality serpentinites and the tholeiites like spilites? What is the density of various greenstone belts? What was their density prior to dehydration during granitoid intrusions?

12. What is the relationship between the greenstone belts, the granite terranes, and the "grey-gneiss" (or Na-granitoid gneiss) complexes? Are there links between the silicic to intermediate volcanic rock sequences *within* the greenstone belts and the surroundings granitoid terranes?

13. Did hydrated simatic rocks of greenstone belts yield the tonalite-trondjhemite granitoids when buried to specific depths? If so, how deep, and was this a tectonic burial (i.e., by subduction or obduction processes) or by continuous volcanic loading?

14. Do granite-greenstone terranes represent successive large-scale tectonic additions around cratonic nuclei? Is there a progressive younging of greenstone belts across shields? What are the age relationships among greenstone belts of different metamorphic grades (where exposed across deep crustal structures such as in the Superior Province, South Indian shield, and the North Kaapvaal craton)?

15. Given a higher Archean radioactive heat production, how was this most efficiently dissipated from the planet? Were Archean mantle temperatures similar to those today (given a greater rate of tectonic processes like spreading) or was the Archean mantle substantially hotter? Did the Earth have a greater total length of spreading ridge or was spreading faster? Was the effective heat loss punctuated, or episodic?

16. Was the average oceanic Archean oceanic crust hotter, younger, and less dense than the average oceanic crust today? If so, how did this affect global-scale tectonics? For example, if there was subduction, was it of shallow-angle type? If the oceanic crust was of very low density, would it have resisted subduction to form interoceanic thrust complexes? Either process could theoretically produce vast amounts of tonalites. Can the geology of the granite-greenstone belt terranes help to distinguish between such processes?

17. Is the abundance of greenstone belts relative to granite-gneiss complexes in Archean terranes geologically representative, or has there been selective preservation of certain rock types and assemblages?

18. What can greenstone belts tell us about the size and distribution of Archean cratons and their lithospheric thickness? Was a thick lithosphere and/or tectosphere the rule or the exception on the Archean Earth? Did Archean continental crust form part of large supercontinents that broke up more frequently than in the Phanerozoic, or was the Archean Earth covered by a multitude of plates carrying continental "icebergs"?

IV. Summaries of Technical Sessions

M. J. de Wit and L. D. Ashwal

The following summaries were prepared from the submitted abstracts and from videotapes of the sessions. We apologize to those whose comments were inaudible, and to those who we may have misinterpreted.

After welcoming comments by LPI Director Kevin Burke and a general outline of the workshop by Chairman Lew Ashwal, Maarten de Wit opened the technical sessions with an introductory talk outlining some of the major questions pertinent to greenstone belt tectonics that he and Ashwal formulated. These 18 questions are outlined in Section III. De Wit attempted to focus the major issues to be dealt with at the workshop by illustrating numerous of these questions with examples from the Barberton greenstone belt of South Africa. In so doing, he introduced question (1), which deals with the possibility that greenstone belts from different cratons may have different features, and represent different tectonic settings. His next point, applicable to questions (2) and (3), concerned the deformational aspects of greenstone belts. He showed slides of two types of large amplitude folds. Both early recumbent and later upright-style folds are associated with major thrust faults and locally with extensive flattening deformation. The Barberton belt can be considered a fold-and-thrust belt with the main tectonic transport having taken place from southeast to northwest. In the northern part of the belt, Barberton was thrust over granitoids, whereas in the south, sedimentation was taking place concurrently. Strike-slip motions are more evident in greenstone belts other than Barberton. The next question (question (6)) related to whether greenstone belts contain ophiolites. He showed spectacular examples of pillow lavas, and pointed out the ambiguity of using them to determine younging directions. Gabbros with cumulate textures are also present, and these have layering orthogonal to that of surrounding mafic-ultramafic layering. This raised the question (question (8)) of the intrusive vs. extrusive nature of komatiites. Slides were also shown of mafic-ultramafics with deformational textures, possibly equivalent to ultramafic tectonites found stratigraphically at the lower most positions in ophiolites. He also showed fine examples of sheeted dike or sill complexes, with both one-way and two-way chilling features. Small chert remnants occur in places between crosscutting intrusives. Regarding the relationship of the felsic volcanics of greenstone belts to surrounding granitoid plutons (question (12)), de Wit pointed out that Barberton dacites and rhyolites have similar light REE-enriched patterns to surrounding tonalites, precluding an origin by simple fractionation from mafic-ultramafic volcanics, which have flat REE patterns. An origin by partial melting of the (hydrated) mafic volcanics is permissible, if garnet is retained as a residual phase. De Wit stated that the model he and coworkers favor for Barberton involves thrusting of simatic crust over a hot, granitoid-plutonic environment, leading to syntectonic granitoid gneisses at the contacts with the greenstones, and silicic magmas emplaced into mafic rocks along thrust zones from which silicic volcanics were erupted subaerially. De Wit next considered the compositions of komatiites (question (10)), showing diagrams with good correlations between MgO and H_2O (which ranges from 2–16%) for Barberton mafic and ultramafic rocks. Compositions of olivines from both spinifex- and cumulate-textured komatiites plot along 100% olivine accumulation lines; de Wit therefore suggested that these rocks have been affected by olivine accumulation and/or major Mg-metasomatism. As a result of this extensive alteration and hydration, the mafic and specifically the ultramafic rocks have very low densities. This observation led on to question (11). De Wit calculates that the bulk density of the Barberton belt, assuming an approximate ratio of spilite and serpentinite of 4:1, is about 2.67 gm/cm^3, essentially equivalent to that of granite. This, he stated, is of importance to those who model the geometry of greenstone belts by assuming density inversions caused by relatively heavy mafic units overlying lighter granitoids. De Wit explained that his model of thrusting of mafic units on top of a terrane undergoing active granitoid intrusion could be accommodated either by (a) low-angle subduction, as proposed by D. Abbott, or (b) interoceanic thrusting, producing tonalitic melts at deeper levels by tectonic stacking, as proposed by V. R. McGregor. Such thrust stacks could produce a stable granite-greenstone terrane that could sustain underthrusting and subduction along its margins. Regarding the size and thickness of Archean cratons (question (18)), de Wit showed five crustal cross-sections from the Kaapvaal craton, which demonstrated that Archean continents contained 30–40-km-thick crust overlain by sedimentary basins (of various tectonic types, including rift and collision related). The presence of old (ca 3.2 Ga) diamonds in later kimberlite pipes implies that these continents were also underlain by thick (150–200 km) lithospheric keels. It is uncertain whether such thick lithosphere is characteristic of all Archean continental nucleii, or are more locally preserved features such as can be inferred below the Kaapvaal craton (questions (18) and (15)). De Wit showed a map of Gondwana, illustrating that, if surface exposures of Archean rocks are assumed to be continuous below younger cover, about 80% of the present continents are pre-2.5 Ga. This, combined with the knowledge of minimal cratonic crustal thicknesses, would favor an Armstrong-type continental growth model, which purports that most of the Earth's present continental mass was extant by 2.5 Ga. Large, early continental masses, combined with the higher expected Archean heat production, leads to the question (also (18)) of the distribution of continental material about the globe. Possibilities include numerous small plates with continental "riders," or a more efficient (higher periodicity) or dispersal of single supercontinents into smaller plates and microplates as seen during the break up of Pangea in the Phanerozoic era. These matters relate to the general question of the temperature of the upper mantle as a function of time (question (15)). De Wit ended his presentation by pointing out a crucial problem in greenstone belt geology: interpretation of field observations (question (5)). He illustrated this by showing slides of well-developed turbidites with Bouma sequences from Barberton. These have been interpreted in opposite ways by experienced American and British sedimentologists, one

favoring shallow-water deposition, and the other a deep-water environment. Clearly, errors of interpretation made at this level can only lead to houses of cards.

The first session on rock components, sources, provenance, and structure of greenstone belts started with two invited keynote talks. Both overviews emphasized the complexities of greenstone belts attributable to frequent structural repetitions by various types of faults and folding. In the first paper, D. R. Lowe and G. R. Byerly focused on the difficulties facing field geologists in interpreting their observations; they argued that many of the current controversies of greenstone belt petrogenesis, sedimentology, tectonics, and evolution arise from an inability to develop a clear stratigraphy in the belts. They illustrated this using the following four types of field analysis that they believe are frequently the cause of misinterpretations, and to which stratigraphic analysis in the early Precambrian greenstone belts is particularly prone: (a) determination of facing directions; (b) correlation of lithologic units; (c) identification of primary lithologies; (d) identification of structural contacts. The absence of fossils has of course long frustrated many attempts to apply classical stratigraphy in greenstone belts; Lowe and Byerly point out that the recent "precise" zircon dating in the Canadian shield will do little to alleviate this. Another important development for greenstone belt stratigraphers is the growing awareness that metasomatic alteration of primary lithologies has, in the past, led to erroneous lithostratigraphic interpretations. Lowe and Byerly pointed to several recent studies that have shown that igneous lithologies of silicic to andesitic compositions are in fact silicified mafic-ultramafic rocks, and therefore that some of the classic ultramafic-felsic volcanic cycles reported from, for example, the Barberton greenstone belt, may not be real. The silicified lithologies reflect complicated and little understood metasomatic processes operative on regional scales. Lowe and Byerly pleaded for new studies that should focus on providing unambiguous criteria to distinguish different stages of alteration and identify their respective environments. Similarly, they emphasized the need for clearer documentation of features indicating structural discontinuities in greenstone belt successions: Until such time that this is done, they feel that the possibility that there are thick stratigraphic sequences in greenstone belts should be retained as a working hypothesis.

In contrast, J. R. Vearncombe et al., in the second overview paper, point out that the geophysical evidence from a number of belts suggests that they are shallow (rarely greater than 10 km, and usually less than 5 km); these are figures considerably less than quoted for the proposed stratigraphic thicknesses of these belts (between 17–45 km as reported from many greenstone belts). This shallow depth suggests to them no simple rotation of the usually upright greenstone belts but, instead, a structural truncation that may be a major decollement zone. They believe this is in accord with the observed greenschist metamorphic facies of the thick successions, which do not display higher grades of metamorphism that would be expected at the base of such thick sequences. Vearncombe suggests this can be explained by involving structural repetitions above a flat, shallow decollement in an imbricate stack with associated folding.

Contrasts and comparisons among various globally distributed greenstone belts and between the igneous components of greenstone belts and Phanerozoic ophiolites were presented in both review talks. Vearncombe emphasized that abundant spinifex-bearing units are restricted to the Archean greenstone successions. Contrary to popular belief, these textures (which indicate rapid crystal growth) form not only in lava flows but also in shallow-level intrusions. They suggest the term "cooling units" be used for spinifex-bearing units and urge identification of new criteria that might permit the environment of their emplacement to be determined more precisely. Major differences in lithological content and proportions among greenstone belts were noted. Lowe and Byerly suggested that a possible implication of the variability among greenstone belts is that they may represent tectonic settings as varied as those represented in modern orogenic belts. Vearncombe noted that various tectonic settings may be represented within individual greenstone belts (e.g., Barberton); he and colleagues believe that this is due either to: (1) progressive evolution of tectonic environments; (2) the superposition of different tectonic environments; or (3) parts or all of the belts being allochthonous, with representatives of tectonically juxtaposed environments. Vearncombe also stressed that there is a paucity of detailed structural observations to show that greenstone belts are synforms, as historically assumed, and furthermore pointed out that the role of granitic diapirism in controlling the structure of greenstone belt successions may have been over-emphasized. There was disagreement between the two talks as to whether or not the older greenstone belts (> 3.0 Ga) are substantiately different from the younger greenstone belts. Lowe and Byerly emphasized a difference. They stated that volcanic sequences in the older belts accumulated under the shallow water conditions of anorogenic platforms, whereas they emphasized that those in the younger belts formed in tectonically active settings. Vearncombe expressed that the idea of such a distinction is tenuous, at least for greenstone belts of the Kaapvaal craton.

At the start of the general discussion D. Lowe clarified his conception of early Archean platform conditions. He emphasized that these platforms are simatic and stable. Only at later stages did tectonism influence sedimentation and further mafic-ultramafic volcanism. Lowe emphasized that quartzites in these greenstone belts are not diagnostic of that early platformal stage, but that they represent some other type of tectonic platform. Both P. Hoffman and K. Burke commented on the extreme complexity of the geology as described in the overview talks, and wondered if too bleak a picture was being painted regarding the prospects of resolving the tectonic history of greenstone belts. Lowe replied that he had not intended to convey a depressing message. He re-emphasized that with further detailed field work many of the controversies could be resolved. M. Bickle pointed out that the silicification and other alterations of rocks reported even in the Barberton greenstone belt were also very marked in the (similar age) greenstone belts of Pilbara, Australia. He then asked if people agree that such alterations were less prevalent in the younger greenstone belts. P. Hoffman replied that the extent of alterations did not show

secular change. He pointed out that in parts of many of the younger Canadian greenstone belts there is extensive alteration. Hoffman felt that Canadian geologists were better equipped to understand the alteration systems because they have succeeded in finding the Cu and Au deposits that were part of these systems. Bickle commented that intense alteration due to tropical weathering in Africa and the Pilbara had obscured the earlier alterations and thus added an additional degree of complexity in these belts. De Wit suggested that like deformation processes, alteration processes take place in various tectonic environments. He pointed out that the alterations documented in the Canadian greenstone belts may have occurred in a totally different tectonic setting from those of Barberton, and that the various styles of metasomatism and associated mineralization may provide valuable insights to their tectonic settings.

Chairman L. Ashwal focused the discussion on the field observations. He asked what could be done about the fact that different field geologists interpret the same features in the same rocks in different ways. He cited "way-up structures" and "sheeted dikes as opposed to lava flows" as examples. G. Harper responded by asking de Wit about the volume of the sheeted dikes in Barberton (shown during his introductory talk and his poster presentation) relative to the total mafic-ultramafic rocks. De Wit replied that this was difficult to judge. The main obstacle was a lack of sufficiently good outcrop of contact relationships to distinguish dikes from flows. He suggested that up to 80% of the Komati Formation might be intrusive. E. Nisbet reiterated that this interpretation hinges on sparse outcrop and he wondered if the intrusives could not have been horizontal in a lava flow setting. De Wit agreed that he could not be completely sure whether the intrusives represented sheeted sills or sheeted dikes, but he continued by saying that other circumstantial evidence led him to believe that they were vertical intrusions. G. Harper stated that from his field experience with Phanerozoic ophiolites, sheeted dikes are not always perpendicular to the bedding of the overlying sediments. Although an orthogonal relationship is commonly assumed, very few detailed studies in ophiolites have actually related the orientation of sheeted dikes to overlying sediments. Neither in the Josephine nor Troodos ophiolites (as recently shown by E. Moores) are these two rock sequences disposed at right angles to each other. Harper believes that the reason for this is that ophiolites form in extensional (tectonically active) environments where characteristically large scale rotations of up to 30°–50° occur. In these environments the dikes may be intruded vertically but are subsequently broken up and rotated by listric normal faults. The sheeted dikes may thus end up with orientations of up to 50° to bedding in the sediments. Harper further suggested that such angular relationships may be further rotated into parallelism during subsequent compressional deformations. What bothered Harper more about the sheeted dikes described from Barberton is that they contained chert xenoliths. He asked how that might be explained. De Wit suggested that they may have been incorporated as a result of polyphase extension in different directions during the emplacement of the vertical intrusions, for which there was clear field evidence.

A. Wilson reiterated the problem of field identifications in greenstone belts, with respect to intrusive rocks within volcanic

sequences. He stated that in Barberton, for example, one must be extremely careful about identifying what the igneous rock units were. The Kromberg Formation, for example, is a sequence of units that has been mapped and documented over and over again as a volcanic sequence. Yet in the lower parts of the type section one can observe a thick serpentinized dunite, then a harzburgite overlain by a gabbro; the textures show unequivocally that these rocks are cumulates, with layering at high angle to the overall stratigraphy.

L. Ashwal steered the discussion to other field-related observations by asking about the criteria used to distinguish thrust faults. De Wit replied that his group had initially relied on the use of opposing facing directions across lithologic contacts. In such cases they infer the presence of faults, whether penetrative deformation fabrics are present or not. He also emphasized that the facing directions determined by his group do not always agree with those determined by D. Lowe in the same area. De Wit explained that other observations also indicate substantial thrust faults. Some of the field data (such as tectonic truncations, rotated unconformities, etc.) could be seen displayed on the posters of S. Lamb and I. Paris. In many places in the study area in the southern parts of the Barberton greenstone belt, Onverwacht-like rocks had been documented to be structurally overlying sediments of the youngest Moodies Group rocks. There are thus clear examples of older rocks lying on top of younger rocks. M. Bickle inquired how the older rocks were dated, and because the answer was by Sm/Nd systematics, he suggested that these dates might be unreliable. However, $^{39}Ar/^{40}Ar$ stepwise heating experiments have confirmed the Sm/Nd ages. F. Schwerdtner pointed out that facing directions do not generally change across thrust zones, unless they are associated with large recumbent folds. He suggested, therefore, that many thrusts may not yet have been picked up in the Barberton. De Wit explained that the reason for starting with an approach of looking for opposing facing directions (and recumbent fold nappes) was to convince themselves that allochthonous units existed within the Barberton stratigraphy. Their philosophy was to prove this first since this assured the presence of thrusts, which are otherwise difficult to prove using only lithostratigraphy. Lowe commented that he disagreed with almost everything de Wit had documented about the facing directions in Barberton. He had looked at the same rocks, but he emphasized that these local controversies could not be resolved at this meeting. Lowe explained, however, that there was no doubt that there were thrust faults in the Barberton terrane. The basic area of disagreement, however, was in which parts of the stratigraphy these thrusts occurred. Whereas there appeared to be no ambiguities about thrusts in the upper parts of the stratigraphy, he believed (unlike de Wit) that there was no evidence for structural repetitions in the lower parts of the Barberton stratigraphy. P. Wyllie asked if de Wit and Lowe had looked together at the same outcrops. The answer was no, and Chairman Ashwal wondered aloud if this might be possible to organize at all, and if so, should they shoot the one that was wrong? W. R. Muehlberger observed that since they would probably still disagree while examining the same outcrop together, they should both be shot. The discussion wisely moved to another, as yet, less controversial topic: are komatiites part

of (Archean) oceanic crust? M. Bickle was dubious to accept this for two reasons. First, the rocks were so tectonized it would be difficult to prove. Second, isotopic data from komatiites such as Kambalda (Yilgarn, Australia), which are also part of a tectonized mafic-ultramafic succession, indicate contamination by sialic crust. These komatiites contain xenocrystic zircons that are about 700 Ma older than their hosts. How could one account for this? It seems that the komatiites must have erupted through older continental crust. De Wit replied that oceans start as rifts, so that early komatiites could conceivably have traversed stretched continental crust, while later komatiites could have formed at mid-oceanic ridges. M. Bickle accepted this possibility, but stressed that geochemical and other tools to identify oceanic crust are complicated, and that we must be sure that the data is correctly interpreted before we use it in models. J. Ludden questioned whether the volume of komatiites in greenstone belts may not be overstated. He pointed out that they are only a minor constituent in the Canadian shield and that they are absent in several greenstone belts. The discussion then moved to the question of the densities of greenstone belts. M. Bickle commented that most rocks in the Barberton are mafic and have densitites of 2.9-3.0 gm/cm^3, and K. Card stated that the Canadian greenstone belts had average densities between 2.85-2.9 gm/cm^3. De Wit replied that unlike the Canadian greenstone belts, he estimated that 25% of the igneous rocks in Barberton were ultramafic rocks, and since these were almost entirely serpentinizated, these rocks lowered the overall density of the Barberton greenstone belt to the granitoid-like values that he had calculated. J. Ludden cautioned, however, that the serpentinization might be a late alteration, in which case the low-density calculations had no relevance to understanding early tectonic processes. De Wit replied that the alteration of the Barberton komatiites almost certainly occurred early in the history of the development greenstone belt (as indicated by old $^{39}Ar/^{40}Ar$ and K-Ar ages) and that this had probably occurred during hydrothermal alteration at spreading ridges where the komatiites were emplaced. E. Nisbet commented that stable isotope measurements from most of the greenstone belts, and in particular those from Southern Africa, indicate that there was both early serpentinization, which was hydrothermal, and late alterations, which were recent. He cautioned de Wit to be very careful with his density estimates. De Wit accepted that. K. Card asked de Wit if he had measured actual densities of the Barberton rocks. De Wit replied no, and explained that the density values chosen for his "back-of-the-envelope" calculations were taken from standard densities given for serpentinites.

K. Burke moved the workshop to a totally new topic. He asked what significance should be attached to the fact that there are different proportions of different rock types in different greenstone belts. He went on to answer his own question and suggested that Archean students might do well to look closer at younger mountain belts, where one finds tremendous variations in rock types along strike. He focused on the example of the ophiolite distribution in the Appalachians: Such complexes occur in abundance in the northern parts of this mountain belt, but were almost absent in the central and southern parts. Burke also emphasized that different tectonic regimes are preserved in younger belts. Thus, because of the fact that in the greenstone belts there is such tremendous diversity, Burke wondered if one should worry about it too much; comparative studies of greenstone belts and younger orogenic belts might be a fruitful approach. After all, Burke continued, greenstone belts are entirely confined to continental crust today. However, they contain thousands of km^3 of tholeiitic pillow lavas, which must be oceanic representatives, and therefore mark places where oceans opened and closed. By analogy to modern belts, we would expect tremendous diversity in what little is preserved in greenstone belts. Burke was therefore unhappy about statements that the old greenstone belts are significantly different from the younger greenstone belts, given that the total area of greenstone belt preserved in proportion to the total area of continents is so small. Burke believes rock preservation is subject to the lottery of extreme tectonism and erosion.

Following the much needed coffee break, McGill and Shrady presented a preliminary structural model for the southwestern part of the Michipicoten greenstone belt, Ontario. Based on their detailed mapping, they have been able to show that this part of the belt is a tectonic complex, consisting of several lithologic packages of clastic sediments bound by early faults. In some cases, the stratigraphic way-up reverses across these faults. On a large scale the boundary faults parallel lithologic layering; on the outcrop scale they commonly cut across the stratigraphy. Only locally do they preserve evidence of penetrative strains. McGill and Shrady demonstrated that the belt contained large recumbent folds, facing in opposing directions, that had developed synchronous with and/or were overprinted by the major faults. They concluded that the simplest explanation for these structural relationships would be a rotated imbricate thrust belt. In the following talk, Hudleston and coworkers presented structural data to show that transcurrent (wrench) faulting in an overall compressive stress regime (transpression) was the cause of the main structural features in the greenstone belt of the Vermilion district, NE Minnesota. Previous work assumed that the deformational features of the belt resulted from pure compression, related to diapiric intrusions of batholithic granitoids to the north and south. Hudleston *et al.* demonstrated, however, that the Vermilion fault was the latest, most brittle expression of a long-term regional dextral shear regime. They concluded that the Vermilion district was a region of relatively soft lithosphere, deformed between two more rigid blocks (either thicker or cooler) to the north and south.

Wrench faulting was also the subject of the next talk by C. Hubert and J. N. Ludden. They presented a tectonic interpretation of the structural elements of the Abitibi greenstone belt (Ontario and Quebec). Hubert and Ludden showed that this belt consists of large-scale lozenge-shaped blocks each with different lithologies and structural-metamorphic histories, separated by faults and zones of ductile deformation. Sedimentary accumulations occur along the margins of the blocks in a series of long, narrow basins bound by the shear zones. Most of the blocks contain an older (2850-2720 Ma) calc-alkaline volcanic-plutonic basement, overprinted by tholeiitic volcanism at about 2700 Ma. Hubert and Ludden propose that the southern Abititi belt formed in a series of rift basins that dissected an earlier formed volcanic arc (analogies were made with Japan, New Zealand, Indonesia, and Central

America), and subsequently were accreted obliquely against a more stable continental margin to the north.

During discussion, R. Fripp asked whether the folds and shears discussed by McGill and Shrady were related, or whether the folds totally predated the shearing. Shrady thought that the shears might have formed during the folding. M. de Wit pointed out that they should then use the term slide for such a shear; he also asked what they knew about the tectonic transport directions. McGill replied that he had reservations about the term slide since it conveyed to him a geomorphic term; they had no idea about the transport directions.

M. de Wit commented that in essence the last two talks described large scale continental melange zones of horrendous structural complexities, and he asked how the authors proposed to set about restoring these features so that one might understand the original basin geometries. Hudleston replied that if there was sufficient outcrop they could measure continuous strain sections and remove these visible effects of shear. He explained that this would have to be tied into the stratigraphic columns. The major zones of structural discontinuities would, however, be very hard to deal with because they were poorly exposed and many may well be missed. K. Burke suspected that this might be a thankless and impossible task; he suggested that they should rather ask questions they could answer more satisfactorily. Trying to correlate stratigraphy among the various blocks in the Southern Abitibi would be difficult indeed. He therefore proposed that they abandon such an exercise (for the moment) and first study more carefully the boundaries between the tectonic blocks, where they are exposed (such as the Larder Lake fault zone). Burke felt that there was a need to learn a lot more about these shear zones; moreover they contain lots of gold and were thus likely to be profitable targets for other reasons also.

F. Schwerdtner told the audience that he had been asked to talk a little bit about the work of G. Stott, who had been mapping in an area just north of that described by Hudleston. The two areas display virtually the same structural styles and histories, but Stott had the advantage of being able to integrate his structural data with detailed geochronology. Stott has devised a scheme in which an originally (D_1) foliated terrane was subsequently subjected to the transpressional strain (D_2). He had also been able to show that some rocks were laid down after the first period of deformation and only experienced the transpressional strains, while other rocks were even younger and undeformed. K. Burke asked what the sort of time frame was for the transpressional strains. F. Corfu, who has worked closely with Stott on these structures, explained the detailed geochronology of the area. In particular he pointed out that a series of deformed volcanics and a deformed pluton were dated at 2689 Ma and 2690 Ma, respectively. In contrast a post-D_2 pluton had been dated at 2684 Ma, so that (given uncertainties) the maximum duration of the transpressional event (D_2) was 10 m.y.

P. Hoffman asked Hudleston whether in his area the transpression was the only deformation or whether it was also superimposed on an earlier event. Hudleston replied that the transpression had been superimposed on a D_1 event that produced neither a metamorphic pattern nor a recognizable strain pattern. He assumed therefore that the D_1 structures

were soft sedimentary structures. Thus, there was no evidence for significant tectonic events prior to transpression. He noted that this was a difference between his area and that of Stott's, and asked therefore what the nature of D_1 was in Stott's area. Schwerdtner replied that he did not know, nor could he answer Hudleston's further question as to whether the pre-D_2 foliation in Stott's area had an original steep or subhorizontal orientation. Some inaudible comments followed until P. Hoffman could be heard to ask if a lot of the granitoid plutons had intruded during transpression. R. Bauer replied indecisively, and P. Hoffman commented that one tectonic setting where granitoid plutons could occur in a zone of transpression was in an arc that had developed over an obliquely subducting slab. M. Jones brought the discussion back to the fundamentals of kinematics. He asked if Hudleston and his coworkers had attempted to use structures such as fibrous extension veins to understand the incremental strain within the shear zones. Hudleston's reply was negative. P. Hoffman questioned the soft sedimentary explanation for D_1 in Hudleston's area, since he had described giant D_1 folds. Hudleston agreed that this was indeed an enigma that he still did not understand.

As a final remark, P. Morgan commented that Hudleston had said that the localization of the transpression had occurred in hotter or thinner crust. Morgan pointed out, however, that as a consequence of the greater strength of olivine compared to quartz, lithosphere with thin crust is stronger than lithosphere with thick crust. Consequently, the transpression zone was localized in hotter or thicker crust. Hudleston agreed and said that all they know is that it was a weaker zone, and they should have referred to lithosphere rather than crust; he said they would revise their abstract.

In the following talk D. R. Lowe and G. R. Byerly discussed the sedimentology and stratigraphy of the southern Barberton belt, South Africa in terms of changing provenance and stability. They have identified three evolutionary stages: (1) A volcanic platform stage during which at least 8 km of volcanics and thin sediments of the Onverwacht group accumulated. The sedimentary units are predominantly shallow-water with little, if any, contribution from uplift and erosion of older basement. These features and the regional stratigraphic continuity are interpreted in terms of a broad, low-relief, anorogenic setting. (2) A transitional stage of developing instability during which distal volcaniclastics and carbonaceous cherts of the Fig Tree Group were deposited. (3) An orogenic stage that involved cessation of active volcanism, extensive thrust faulting, and widespread deposition of clastic sediments of the Moodies Group. The overall sequence includes many local unconformities, and at least one major break between the Fig Tree and Moodies Groups. This break is interpreted as an unconformity complicated by later structural movement.

A companion paper by Byerly and Lowe dealt with Barberton volcanism. In their area they estimate a minimum thickness of 12 km for mafic and ultramafic rocks of the Onverwacht Group, and 1 km of volcaniclastic sediments of the Fig Tree Group. Structural repetitions in the Fig Tree Group resulted in a greater tectonic thickness. Some units, previously interpreted as polycyclic volcanic successions, appear to contain thick alteration zones where mafic flows have been subaerially silicified. Komatiites are mainly spinifex-textured flows, with

evidence for crystal accumulation near their bases. Basalts include pillowed and tuffaceous units as well as sills and dikes, and may represent near-vent facies similar to modern cinder cone fields. Dacitic volcanics commonly include tuffaceous units and thick flows, pyroclastics and epiclastics, some of which can be interpreted as vent complexes. The komatiites and basalts could be related by low-pressure olivine fractionation, but there are insufficient cumulates present *in situ* to account for this. Dacites are consistent with having been derived by small degrees of partial melting of a mafic, amphibole-rich source.

In discussion, J. E. Martyn asked Byerly if the K and Rb enrichments observed in his proposed alteration of basalt to silicic rocks might be better interpreted as having taken place subaqueously, for example, by hydrothermal alteration rather than by subaerial weathering. Byerly responded that the alteration certainly could be hydrothermal, considering the evidence for extensive igneous activity, but in this case the occurrence of these altered rocks as large regional stratiform zones would require an unusual flat-lying style of water circulation. Such models have been proposed by de Wit for Barberton and by others for Pilbara rocks. Byerly also pointed out that K-enrichment of basalt can occur on the ocean floor of 5–10 Ma, without the high temperatures associated with hydrothermal activity. Byerly explained that their evidence for subaerial weathering comes from associated sedimentary rocks.

M. de Wit commented extensively on both Lowe and Byerly's talks. He stated that he was in agreement with many of Lowe and Byerly's points, specifically with regard to the dacites and the silicification of the mafic and ultramafic rocks. He had in fact proposed in the early 1980s that many of the dacites might have been differentiates of the mafic rocks, but this is clearly not the case. However, he strongly disagrees with Lowe and Byerly's interpretation of the total stratigraphy of the Barberton rocks. De Wit explained that he and his colleagues approach the Barberton belt in terms of a much more complicated history, involving thrust nappes and major allochthonous relationships, and therefore have largely abandoned the previously established stratigraphy. He feels that some of the features interpreted as weathering zones by Lowe and Byerly are tectonic shear zones developed very early during the igneous history of the Onverwacht Group. He showed a slide of one of these shear zones that consists almost entirely of extension veins separated by mylonites and schistose material, which is often fuchsitic. A second slide showed the polyphase tectonic history of these shear zones with folding of the extension veins. He offered participants an opportunity to examine these samples at his poster display. He also illustrated, with slides of maps from the Komati and Kromberg type sections, his point that some of the rocks interpreted by others as conformable extrusives are actually intrusives, crosscutting earlier units. These include spinifex-textured komatiitic rocks as well as dacites that he interprets as intrusive domes. Some of the shear zones he described are actually caught up as megaxenoliths in the mafic-ultramafic intrusive rocks, arguing for an early active (extensional) tectonic environment rather than a simple platformal style of deposition for the Barberton rocks. He explained that there is still major uncertainty as to whether this took place in a shallow water environment similar to Iceland, or in a deeper water, mid-oceanic ridge-type setting. He showed several maps illustrating the numerous major thrust zones he and colleagues have identified, and stated that although they are uncertain as to exactly how to restore these units stratigraphically, the total thickness of the mafic-ultramafic rocks cannot be much greater than about 3 km, according to their interpretation.

In response, Lowe explained that the final stratigraphy he and his colleagues adopted was based on careful consideration of earlier interpretations, such as those of Anhaeusser and the Viljoen's, which involve large-scale continuity, as well as de Wit's, which involves abundant fault-related disruptions. He agreed that the fibrous veinlets de Wit showed formed under vertical tensional stress, but stated that the origin of that stress is very ambiguous. It does not necessarily imply any slip along those zones. Similar veinlets can be found in purely sedimentary environments, such as in gypsum or satin spar horizons or in ice crystallized in permafrost zones. Regarding continuity of mafic-ultramafic flow units, Lowe admitted that not every flow is traceable over large distances. Individual flows are often lenticular, and there are other complex facies changes. Accordingly, they have attempted to recognize *groups* of volcanics, tens to hundreds of meters thick, that *are* traceable over larger distances. Examples include ultramafic zones and pillowed tholeiitic sequences in the Hooggenoeg Formation. Lowe also agrees that intrusive komatiites are abundant, but this says nothing about the presence or absence of *extrusive* komatiites. He pointed out that it would be staggering to have hundreds of meters of extrusives in the Fig Tree and none in the Onverwacht. His evidence for abundant extrusive komatiites in the Fig Tree includes the presence of fragments of komatiite and detrital Cr-spinel in the bases of overlying chert units. There are also cracks and brecciated zones filled with sedimentary material penetrating from cherty units into underlying komatiites. Therefore, although there is good evidence for both extrusive and intrusive komatiitic rocks, the relative abundances are uncertain. Finally, Lowe agreed that a fault clearly exists in the felsic volcanics of the Hooggenoeg Formation, but commented that it is possible in many places to trace stratigraphic markers across it, including komatiites, and a newly recognizable evaporitic unit.

Discussion continued with comments by R. A. Hart who apologized for being a geochemist; he was convinced early by his brother Stan that geochemistry would soon render field geology obsolete. Hart showed slides documenting that K-Ar ages demonstrate that the last metamorphism of Barberton rocks, including serpentinization and Si-metasomatism, took place 3.3 Ga ago. Oxygen isotope data on ultramafic tectonites indicate that they formed in contact with fluids similar to sea water, favoring a submarine rather than a subaerial site for alteration. Similarly, Barberton cherts have relatively low ^{18}O compared to biogenic cherts formed on the present-day ocean floor, suggesting that they are also hydrothermal in origin. Hart favors an ocean-ridge environment for the hydrothermal activity, which is consistent with the ophiolite model for this terrane. J. N. Ludden asked Hart why there are no comparable massive sulfide deposits in Barberton as in the Abitibi belt of Ontario, if the hydrothermal circulation model is correct. De Wit responded that Barberton contains abundant Au mineralization, and that the apparent absence of Pb-Zn deposits may be an

indication of differences in tectonic styles between Barberton and Canadian greenstone belts, for example, mid-oceanic ridges rather than island arcs. R. E. P. Fripp commented that his detailed mapping in the Theespruit area of Barberton shows at least seven major tectonic zones, and he also commented on the possibility of multiple tectonic breaks in the Komati Formation and cast his vote for 3 km rather than 8–12 km of stratigraphic thickness of the Onverwacht sequence. P. F. Hoffman asked whether the extension faults that were mentioned could be related to caldera collapse. Byerly responded that there is no evidence for that. The faults are developed on top of felsic volcanic edifices, but there are no associations that indicate eruptive activity post-dating faulting.

Chairman Lew Ashwal adjourned the controversial morning session with regret that discussion must come to an end.

The afternoon session opened with two talks drawing analogies between Archean greenstone successions and Phanerozoic ophiolites. G. D. Harper described a (2.7 Ga?) sequence of "ophiolitic" rocks, including pillow lavas, metadiabases, metagabbros, and ultramafics from the southeastern Wind River Mountains, Wyoming. The rock assemblage has been multiply deformed and metamorphosed, and consists of a series of tectonic "slices." Numerous primary textures have been preserved, including metadiabases containing parallel dikes with one-way chilling features, interpreted as a deformed sheeted dike complex. Ultramafic rocks (serpentinites) and associated metagabbros have relict cumulus textures. Associated sedimentary rocks include pelites, quartzites, and banded iron formations. Immobile trace elements are similar in pillow lavas and metadiabases, suggesting that they are cogenetic. All the units of a complete ophiolite are present except for mantle peridotites.

H. Helmstaedt and W. A. Padgham described an approximately 11-km-thick sequence of dominantly mafic metaigneous rocks from the Yellowknife greenstone belt in the Slave Province of the Canadian Shield. The mafic rocks grade upwards from coarse-grained locally layered gabbro, through a multiple dike complex, into massive and pillowed flows with interflow sediments. This assemblage was compared to Mesozoic ophiolites like the Rocas Verdes of southern Chile, interpreted to have formed in an arc-related marginal basin setting.

In discussion, J. N. Ludden pointed out the similarity of this Yellowknife section to those of modern oceanic islands like the Canary Islands. M. Bickle questioned how far down in the sections the sedimentary horizons appear. Helmstaedt responded that cherts first appear 5 km above the base of the section at Yellowknife, and although iron formations appear at the very bottom, it is unclear if these are in stratigraphic continuity with overlying gabbros or if there is an intervening structural break. W. S. F. Kidd wondered if anyone would object to calling this sequence an ophiolite if there was a basal thrust zone. E. G. Nisbet felt that we must be extremely careful in interpreting these fragmented units as representing ophiolites, pointing out that nearly all Archean basaltic rocks would plot in the ocean-floor field on a Pearce-Cann geochemical discrimination diagram. Harper assured us that he *has* been careful in the Wind River section, and that his combination of detailed structural and geochemical studies are at least consistent with an ophiolitic interpretation. He emphasized that if this was a terrane in the Appalachians or Cordillera, no one would object to calling it an ophiolite. Harper also pointed out that identification of the precise sites of origin of these Archean sequences is less important, at present, than recognizing their possible similarities with Phanerozoic ophiolites. Helmstaedt stressed the recurring semantic problem with placing too much emphasis on the definitions of "ophiolites" and "greenstone belts." Harper reminded us of the entrenched dogma that equates all Phanerozoic ophiolites to ocean floor; some, such as Troodos, are reliably interpreted as having arc affinities. To equate with confidence any given potential ophiolite sequence to an ocean floor environment requires identification of ultramafic tectonites; these have been looked for but not found in the Wind River area. R. Hart asked if any serpentinites had been found, and Harper said that the ultramafics of the Wind River Mountains have relict cumulate textures, which could not allow an interpretation as upper mantle peridotites. P. F. Hoffman voiced a concern similar to Nisbet's about interpreting these sections as oceanic crust formed at specific sites such as spreading centers, ocean islands, plateaus, or island arcs. He also wondered how komatiites fit into the picture. Harper pointed out that not a single Phanerozoic ophiolite (with the possible exception of Bay of Islands) can be demonstrated to have formed at a spreading center. J. F. Casey concurred and stressed that there is little agreement even in Phanerozoic ophiolites as to their sites of formation. Bickle mentioned that he was struck by the relative absence in these Archean occurrences of gabbros, especially ones cut by dikes, leading to his interpretation of their formation on continental crust. Harper stated that in the Wind Rivers, only the coarse-grained lineated gabbros in the upper parts of the section are crosscut by dikes; they are absent from the cumulate-textured gabbros associated with ultramafics. A similar relationship is observed in Phanerozoic ophiolites. Ludden commented that the characteristic weathering profiles and associated sedimentary rocks expected from such ophiolitic assemblages should be easily identified. Harper pointed to the presence of banded iron formations, and Helmsteadt reiterated the presence of sediments at Yellowknife, where there is also extensive alteration in places. The alteration in the Yellowknife belt, however, may be much younger, possibly associated with the emplacement of quartz porphyry dikes. Casey pointed out two differences between Phanerozoic ophiolites and the Archean sequences described here: the presence (in Phanerozoic examples) of ultramafic tectonites, and the absence of interstratified sedimentary horizons within pillowed units. Harper responded that interstratified sediments were absent from Wind River pillow lavas, and Helmstaedt reminded us that sediments were absent from the first 5 km of the Yellowknife volcanic-intrusive section. The absence of ultramafic tectonites was attributed by Helmstaedt to tectonism. M. de Wit emphasized D. Elthon's work on an ophiolite in the southern Chilean Andes, which, although it lacks ultramafics, is still comfortably referred to by most as an ophiolite. De Wit stressed that the important thing is to recognize the Archean sequences as "ophiolitic" regardless of whether or not an ocean floor setting can be demonstrated. Elthon commented that the southern Chile example may lack ultramafics simply because it has not been sufficiently uplifted.

This discussion was followed by two geochemical papers on the Michipicoten greenstone belt and adjacent terranes of northwestern Ontario (Superior Province). E. Smith, M.

Tatsumoto, and R. M. Farquhar presented the results of their combined U-Th-Pb and Lu-Hf isotopic study of zircons from supracrustal rocks of the Michipicoten belt and gneisses of the adjacent Kapuskasing granulite zone, which is thought to represent the underlying basement. The zircons show distinct patterns in concentrations of Lu, Hf, U, and Th, which can be correlated with rock type. Granitic zircons are enriched in U and Th relative to intermediate granitoids, and those from high-grade mafic gneisses show evidence of Lu, Hf, and Pb loss, presumably during granulite metamorphism. Hafnium isotopic ratios of these zircons indicate derivation from three distinct sources: (1) a high Lu/Hf source, interpreted as depleted mantle, that gave rise to tholeiites, (2) a moderately enriched Lu/Hf source, interpreted as lower crust, that produced dacites and subvolcanic equivalents, and (3) a subchondritic Lu/Hf source, interpreted as upper crust, that generated rhyolites and post-tectonic potassic granitoids. The data were interpreted in terms of a model involving progressively shallowing depths of melting, and the close correspondence of ages (2748-2714 Ma) suggests that mantle depletion, crustal extraction, and intracrustal differentiation occurred as part of the same episodic event.

In discussion, Ludden questioned how the isotopic signature of their proposed lower crustal source could be distinguished from mixing between upper crustal and mantle sources. Smith responded that such a case should be expected to result in a continuous range of isotopic compositions instead of separate vectors on a plot of age vs. Hf.

P. J. Sylvester, K. Attoh, and K. J. Schulz interpreted their geochemical data for rhyolitic rocks from the Michipicoten greenstone belt in terms of a convergent plate margin model rather than one involving intracontinental rifting. Michipicoten rhyolites have high (Al + Ca)/(Fe + Na + K) and a range of SiO_2, more similar to Cenozoic subduction-related rhyolites than to those found among intracontinental rifts. These authors drew an analogy between the tectonic settings of the Michipicoten belt and the Taupo volcanic zone, where pyroclastic rhyolites derived from the continental crust of New Zealand are deposited on the adjacent sea floor among tholeiites from the Tonga-Kermadec island arc. Such an environment could possibly account for the mafic-to-felsic cycles of the Michipicoten belt, although the presence there of subaerial and shallow-water volcanics, volcaniclastics, and sediments seemingly requires either intermittent emergence of the volcanic pile, or the presence of small, underlying continental blocks.

J. E. Martyn questioned the analogue between Michipicoten and Taupo, citing the absence in the Archean example of extensive epithermal Au and porphry-related mineralization. He also stated that no thick sequences of tholeiites and komatiites exist at Taupo, and therefore the basement for the rhyolitic rocks at Taupo and Michipicoten was different. Sylvester responded that Michipicoten may be more analogous to the terrane *adjacent* to the Taupo volcanic zone. Ludden pointed out that the Michipicoten belt can be correlated (across the Kapuskasing Structural Zone) with the northern (Chibougamou) part of the Abitibi greenstone belt, but not with the southern part, where there are abundant mafic and ultramafic rocks. This terrane may, therefore, consist of a series of accreted blocks of oceanic affinity. C. H. Shrady commented that

ultramafics are present in the parts of the Michipicoten belt she has worked in, and that most of Sylvester and colleagues' geochemical comparisons were carried out for lower cycle Michipicoten volcanics. Sylvester stated that upper-cycle rhyolites, produced some 50 Ma later, are quite similar geochemically. The ultramafics Shrady referred to are evidently intrusives, not komatiites. W. R. Muehlberger and J. F. Casey wondered if back-arc or fore-arc accretionary sediments could be identified at Michipicoten, and Sylvester stated that these possibilities could not be ruled out. W. S. F. Kidd questioned the interpretation of cyclic volcanicity in the Michipicoten belt, citing McGill and Shrady's morning presentation, which documented evidence for early large-scale detachments. K. J. Schulz pointed out that there is some geochronology supporting volcanic cyclicity, but whether this will stand up in detail remains to be seen.

The session continued with two talks favoring an intracontinental rift environment for certain greenstone belts in the Superior Province of the Canadian Shield. G. R. Edwards and D. W. Davis reported U-Pb geochronologic results for volcanic and plutonic rocks from the western Wabigoon Subprovince, Ontario. Here the volcanic sequence consists of bimodal Mg-tholeiite and rhyodacite (2755 Ma), followed by Fe-tholeiite and rhyodacite accompanied by mafic and tonalite plutons and felsic calc-alkalic volcanics (2734-2718 Ma), and finally dacite (2711 Ma). This volcanic-plutonic terrane is adjacent to a dominantly plutonic terrane (2720-2725 Ma) consisting of domes of gneissic tonalite to granodiorite, and later diorite to granitic plutons. Some gneisses and supracrustal rocks in the latter terrane have ages as old as 3.0 Ga. These relationships are interpreted in terms of rifting of the 3.0 (and older) basement, starting with mafic magmatism, which evolved to bimodal basalt and rhyodacite. Tonalite intrusions accompanied the volcanism. Deformation of the rift sequence and basement complex is attributed to heating of the lower crust by ponded mafic magma.

The geology, stratigraphy, and geochemistry of the La Grande greenstone belt of Quebec was discussed in a talk by T. Skulski. The western part of the belt consists of immature clastic sediments and mafic volcanoclastics, overlain by pillowed and massive basalts; in places this sequence continues with coarse clastic sediments, banded iron formations, and finally andesitic volanoclastics intercalated with immature clastic sediments. In the eastern part of the belt, felsic volcanoclastics are overlain by pillowed basalts, and eventually, komatiites. Skulski and coworkers suggest that the La Grande belt formed on continental crust because: (1) the provenance of clastic sediments changes upward from both intra- and extrabasinal to uniquely extrabasinal, (2) inclusions of metasediment and granitoid occur in volcanic rocks, and (3) they have identified a possible unconformity between supracrustals and older tonalitic basement. They believe the crust initially acted as a density filter to produce basaltic fractionates of komatiitic liquids. Later, the crustal barrier failed, allowing komatiitic magmas to ascent to the surface. Their model, ostensibly consistent with the distribution of rock types and paleocurrent directions in clastic sediments, involves intracontinental rifting, which propagated from east to west across the belt.

These talks stimulated lively discussion, initially led by K. Burke, who felt that rifting models for greenstone belts are

inconsistent with the abundant evidence for large-scale compressional features documented earlier in the workshop. Skulski stated that he and coauthors have focused on the features produced during the early extensional phase, and agreed that the compressional features may have formed during subsequent ocean closure. Burke accepted this. Edwards, however, argued that the Wabigoon Subprovince never developed into a full-scale ocean, and that there was no large-scale compression associated with closure. He also argued that the compressional features present may not have formed under a regional compressional regime, an idea supported by W. M. Schwerdtner. Burke did not seem convinced.

The final two talks of the afternoon session emphasized compressional tectonic features of greenstone belts. J. E. Martyn discussed evidence for structural stacking and repetition of volcanic-sedimentary units in the Norseman-Wiluna greenstone belt (Kalgoorlie District), western Australia, which contrasts with previous models involving polycyclic stratigraphic repetition. Martyn cites several examples where mafic-ultramafic units, originally mapped as stratigraphically separate entities, can be shown to be equivalent when considered in a regional sense. Detailed examination of contacts commonly reveals evidence for strong shearing and cataclasis, and in places, reverse facing directions. Martyn argues that most repetition has been produced by thrust faulting rather than by isoclinal or recumbent folding. The lack of evidence for high-pressure metamorphism or juxtaposition of strongly contrasting domains suggests that thrusting occurred on a scale smaller than that of many Phanerozoic convergent plate boundaries. Martyn favors an intracratonic setting rather than an open plate margin, with intrabasinal gravity gliding resulting from vertical uplift. The instabilities may have been triggered by granitic intrusions. The illusion of a polycyclic stratigraphy may have been caused by later folding and faulting.

P. Hoffman presented his tectonic model for the 2.7–2.5 Ga granite-greenstone terrane of the Slave Province, Northwestern Canada. The greenstone belts are interpreted as remnants of tectonically accreted seamounts, arcs, aseismic ridges, submarine plateaus, and microcontinents. These bathymetric highs, overlain by chemical sediments and flanked by clastics, were buried by thick sequences of orogenic turbidites upon entry into trenches, where they were detached from underlying oceanic units. The latter were subducted. The accretionary complexes were later intruded by post-tectonic plutons. Hoffman believes that this model accounts better for the sequence of volcanic, sedimentary, deformational, metamorphic, and plutonic events than previous theories involving intracontinental rifting. Hoffman's hypothesis may be tested by regional Sm-Nd studies of late plutons, which should yield model ages equivalent to or younger than the volcanic rocks.

K. Burke opened the discussion by asking about events to the east of Hoffman's area. Hoffman explained that the eastern margin of the Slave Province underwent a Himalayan-type collision at about 1895 Ma, presently represented by a suture corresponding to the Thelon Front. Contemporaneously, the Wopmay orogen on the west side of the Slave Province was an Andean-type arc, undergoing episodic extension, shortening, and strike-slip movements. H. Helmstaedt inquired about the absence of rift-related rocks in the Slave Province, and also

about the possible locations of obducted sequences. Hoffman reiterated that no rift sequences are present, and that the greenstone belts represent sheared-off and accreted high-standing oceanic features in a setting analagous to modern-day Japan. Burke attempted to clarify Helmstaedt's question as to why ocean-floor rocks are absent. Hoffman stated that the presence of the upper parts of ophiolitic sequences cannot be ruled out in the Slave Province, but he expressed doubt about such an interpretation for the Yellowknife belt. G. D. Harper commented about the danger of interpreting geochemical data in terms of tectonic environment. Hoffman stated that most greenstone belts in the Slave Province are dominantly composed of andesites, dacites, and high-Al, low-Ti basalts, but at any rate his tectonic model does not stand or fall on the basis of geochemistry. W. M. Schwerdtner asked Hoffman to speculate on how his model would apply to the Superior Province. Hoffman responded by stating that there are a wide variety of features on the ocean floor that could get accreted: oceanic plateaus related to hot spots, fracture zones, etc., so it is entirely conceivable that greenstone belts differing in lithology and chemistry could be generated in different areas. G. R. Edwards commented that if greenstone belts are related to convergent plate boundaries, the absence of spreading features could be accounted for by subduction, but translational features should be preserved. Hoffman agreed, and pointed out that this could be easily accommodated in his model: He attributed the absence of strike-slip features in the Slave Province to orthogonal subduction; oblique subduction could and should preserve translational features. M. Bickle asked Hoffman how structurally complex the turbidites were in Slave, and we learned that they were very deformed indeed. J. R. Vearncombe offered his general impression about the structural papers presented in this session: that greenstone belts of the Superior Province are characterized by abundant strike-slip movements, those of South Africa by thrust faulting, and those of Western Australia by imbricate stacking. He wondered if this is a real difference in style rather than one of interpretation. J. E. Martyn commented that evidence for abundant strike-slip motion is available in Australian greenstone belts also, and he doubted if there were differences in styles between different cratons. D. De Paor recommended that tools used in Phanerozoic structural geology, such as branch-line mapping techniques, be applied to Archean terranes. He illustrated his points with an example from Barberton, where he said mylonites "stare you in the face." Martyn reiterated that we must consider an entire province before making conclusions about tectonic settings of individual greenstone belts. A. K. Gibbs reminded us not to forget that many greenstone belts are Proterozoic, including some of those in South America, West Africa, Southwestern United States, Baltic Shield, Egypt, Saudi Arabia, and Canada. He also commented that while the accretionary events were taking place in Hoffman's area, a classical rift sequence was developing in the Carajas district of Brazil. Hoffman added that the Proterozoic greenstone belts of Manitoba were long thought to be Archean. The session was adjourned by Chairman W. S. F. Kidd.

Session II (Greenstone Belt Externalities: Magmas, Heat Flow, Fluids, and Strain) opened with an invited keynote talk by M. Bickle and E. G. Nisbet on thermal constraints on

greenstone belt tectonics, particularly those available from thermal modeling. The thermal history of the Earth may be calculated from the present distribution of temperature and heat-producing elements by deriving relationships between internal temperature and heat loss. One must also assume that the Earth was hot after accretion, that the present radiogenic heat production accounts for about half of the total heat loss, and that convective heat loss processes varied only in rate in the past. The main conclusions derived from parameterized convection models, which treat uncertainties in convection of fluids with temperature-sensitive, non-Newtonian viscosity, are: (1) Interior temperatures have not changed by more than a few hundred degrees over most of Earth's history, and (2) higher internal temperatures result in thinner plates with higher thermal gradients. Further constraints must come from the Archean rock record, which could provide clues to mantle temperatures and lithospheric thermal gradients. Although the presence of komatiites in Archean greenstone belts has been taken as evidence supporting hotter Archean upper mantle, there are major uncertainties as to the MgO content of primary liquids (and hence eruption temperatures), the degree of melting involved, and the relationship between eruption temperatures and mantle temperatures. Even the most conservative estimates (a 25% MgO, 1500°C lava) imply mantle temperatures about 200°C hotter than at present. The widespread 8–10 kbar, 700°– 900°C metamorphic conditions in Archean gneiss terranes imply crustal geothermal gradients not unlike those of today, and comparable high-P, low-T gradients are also recorded in some lower grade Archean terranes. These relatively low gradients and the occurrence of Archean-age diamonds imply thick, cool lithosphere, seemingly inconsistent with higher Archean heat flux. Stabilization of a similar or thicker Archean lithosphere therefore is a major problem. Bickle suggested one possibility involving density decrease of mantle by melt removal, which occurred to much greater depths in the Archean due to hotter mantle temperatures. All these considerations assume a mechanism of heat loss for the Archean similar to that of today, i.e., plate tectonics. Other mechanisms, such as vertical recycling rather than horizontal motions, are possible, but are inconsistent with the growing geological data base, discussed during this workshop, which suggests major horizontal plate motions during the Archean.

There was extensive discussion of this paper, opened by D. Abbott, who commented that the Archean ages of diamonds should not necessarily be taken as evidence that Archean and modern-day lithospheres had similar thicknesses. T. Jordan's present-day tectosphere is about 400 km thick, and modeling of Archean sedimentary basins yields estimates of about 200 km for ancient continental lithosphere. Bickle pointed out that this depends on which sedimentary basins are modeled, and stated his personal belief that the best constraint on present-day lithospheric thickness comes from estimates of reduced heat flow to continents, which indicates that modern continental lithosphere is about 150 km thick. Abbott asked if any of the convection models Bickle discussed actually have smaller-scale upwelling and downwelling cells, and if he could speculate on how the results would change if these were taken into account. Bickle responded that F. Richter's parameterized models include both plate-scale and smaller-scale convection, but plate-scale

convection is extremely difficult to parameterize because we do not have a proper understanding of the resistive forces of plate tectonics, particularly near-surface ones. Variable viscosity models are so cumbersome that they are not really run over anything like a realistic set of conditions. Bickle felt that we will not get much further until they can be parameterized, for example, by using supercomputers to appropriately scale viscosity against temperature. Even then, however, we will be limited by our inadequate knowledge of the viscosity variation in the mantle. Abbott worried further that the variable viscosity convection models may not be sufficiently realistic. Bickle admitted to the possibility that the results may be off by a factor of two. P. F. Hoffman wondered about Bickle's point regarding the more "depleted" nature of Archean lithospheric mantle, stating that there is some evidence from the North American craton that the Archean lithosphere behaved more buoyantly than that of the Phanerozoic. Bickle then presented the results of some calculations that show that "depleted" lithosphere may actually be slightly more dense than underlying undepleted mantle. This effect is quite subtle, and may allow subduction of old lithospheric plates while still preserving them fairly thick. H. Helmstaedt pointed out that in addition to diamonds, there are other mineral assemblages in xenoliths (which give Archean Sm-Nd ages) that could be used to constrain Archean lithospheric pressures and temperatures.

P. J. Wyllie directed the discussion to the idea (mentioned in Bickle and Nisbet's abstract) of the high-pressure density inversion of komatiitic melt at depth, and the implications for a subsurface magma ocean. Wyllie showed a slide with some recent data of Rigdon, Stolper, and Ahrens, who experimentally shocked silicate liquids to 250 kbar, and confirmed previous calculations that at no great depths, melt densities exceed that of the mantle. This could imply a deep layer of MgO-rich melt (which only rarely reaches the surface) giving rise to and underlying depleted lithosphere. Bickle added, however, that Rigdon has recently found pure diopside to be significantly less compressible than the synthetic mixtures of anorthite/diopside used in the earlier experiments, and consequently Rigdon considers the density inversion of komatiite at depth to be an open question; moreover, as Bickle pointed out, there are many problems with the results from opposed-anvil experiments. Bickle also pointed out that if a subsurface "magma ocean" is to be stable, the solidus curve must slope more steeply than the adiabat at that depth, or else the melt will eat its way through its cap. Once again, experimental evidence on this question is ambiguous, although it remains a possibility. At any rate, no one has adequately addressed the importance of an Archean subsurface magma ocean on the thermal evolution of the Earth. K. Burke and E. G. Nisbet mentioned that both D. Walker and C. Scarfe are reported to have independent experimental support for the komatiite density inversion.

The session continued with three contributed talks on various aspects of Archean melts. P. C. Thurston and L. D. Ayres discussed volcanological constraints on Archean tectonics. They suggest that mafic volcanic sequences in greenstone belts were erupted by sheet flow processes rather than from shield volcanoes, and that felsic units were produced mainly from Plinian rather than Vulcanian eruption types. Both eruption rates and lifespans of Archean volcanic systems are purported to

have been higher than Phanerozoic ones. Volcanological and geochemical data suggest that basaltic melts ponded in sialic crust, where they fractionated and induced crustal melting. They presented evidence that large-scale volcanism-related subsidence kept pace with accumulation. Based on these considerations and the abundant bimodal compositions, Thurston and Ayres favored a rift analogue rather than an island- or back-arc setting for many greenstone belts.

D. Elthon discussed the petrogenesis of komatiites in terms of the compositions of their parental magmas and the degrees of partial melting required to produce them. He reevaluated previous estimates of the compositions of primary komatiite magmas, which contain as much as 33% MgO. Elthon computed the MgO contents of liquids in equilibrium with the most Mg-rich olivine compositions reported from natural komatiites, using experimentally determined Fe-Mg K_D values of 0.28-0.31, appropriate for 1 atm crystallization between 1450°-1650°C. The results indicate primary liquids with much lower MgO (22-25%), implying liquidus temperatures of about 1500°C. Since these olivines do not appear to be in equilibrium with liquids compositionally equivalent to their whole rocks, Elthon concludes that the rocks were enriched in MgO by olivine accumulation and/or Mg-metasomatism. Elthon also used pseudo-liquidus phase equilibria to demonstrate that only at small to moderate increments of melting (< 30%) will melts crystallize augite as the first pyroxene, as observed in natural komatiites. His results are substantially different from previous estimates of high eruption temperatures (> 1600°C) and large degrees of partial melting (50-80%) for komatiites.

W. C. Phinney, D. A. Morrison, and D. E. Maczuga discussed the common occurrence in many Archean cratons of large (up to 20 cm), equidimensional, calcic (An_{80-90}) plagioclase megacrysts, which occur as: (1) segregations of anorthosite or in basaltic intrusives and extrusives associated with greenstone belts, (2) megacrysts in basaltic dike swarms in stable cratons, and (3) anorthosite complexes associated with marbles and quartzites in high-grade gneiss terranes. Attempts to determine parental melt compositions of the megacrysts are hampered by alteration and metamorphic effects. The data at present require isothermal crystallization of megacrysts from tholeiitic melts, followed by entrainment of the megacrysts as their host melts ascended to the surface. These occurrences also seem to require that large volumes of mafic melt underwent similar crystallization histories in both oceanic and cratonic settings.

Nearly all of the discussion in this period related to komatiite genesis. K. J. Schulz asked Elthon if his calculations were based on compositions taken from spinifex-textured olivines; these may not represent the products of equilibrium conditions. Elthon responded that the olivines in these rocks were from both textural types and were assumed to represent equilibrium conditions regardless of their texture. A. H. Wilson asked how Elthon knew that olivine-spinel reequilibration took place in these komatiites. This certainly occurs in layered intrusions where spinels are commonly zoned, and their compositions are dependent on their size and ratio of surface area to mass. He also pointed out that this reequilibration only takes place down temperature, and that although the change in K_D is very small from 1600° to 1000°C, it changes dramatically at lower

temperatures. Elthon stated that the temperatures indicated by olivine-spinel equilibria in komatiites are between about 700°-800°C. Some volcanic rocks that have equilibration temperatures of 1200°-1400°C may have undergone minimal if any subsolidus reequilibration. One problem with komatiite data is that it is often difficult to determine from published reports the proximity of a given spinel analysis to a presumed coexisting olivine. We obviously need more quantitative data with those petrographic observations. Elthon also pointed out that this reequilibration effect is rather small—the MgO contents of inferred parental liquids would only change by a factor of 1-1.5%. E. G. Nisbet stated that analyses of the cores of zoned olivines should eliminate reequilibration effects. Elthon agreed, but stated that diffusion distances for Fe-Mg exchange can be quite large (up to 0.5-1.0 cm for some cumulate rocks). Wilson commented that the cores of zoned olivines are the best ones to analyze, because olivine diffusion rates are so high that even during modest cooling, they commonly become compositionally homogeneous.

Bickle made numerous comments about Elthon's komatiite talk. He stated that Elthon's arguments about komatiite crystallization sequences depend on the choice of mantle composition. Elthon disagreed, arguing that virtually all proposed mantle compositions plot below the join connecting olivine and a low pressure melt. Bickle commented further that olivine compositions are likely to continually exchange with liquids during crystallization, and that Fe-Mg diffusion coefficients are so high that the most Mg-rich olivines may not be preserved. In addition, published olivine compositions represent an appallingly small data set—nearly all analyses come from rocks with about 27% MgO, so the lack of calculated liquids with MgO greater than 30% is not too surprising. Bickle also stated that on his (Bickle's) compositional data plot of komatiitic rocks, a cationic Fe/Mg = 0.23 corresponds to an MgO content of 27% rather than Elthon's 25%. Elthon established, however, that Bickle plotted total Fe instead of separating FeO and Fe_2O_3, which will inherently result in liquids with higher calculated MgO. Bickle also argued that there are enormous uncertainties in K_D over the range of 27-33% MgO. Elthon disagreed, stating that there is a clustering of determined K_Ds at the values Elthon used, and those for high-pressure olivine-liquid equilibria that Bickle used are probably not appropriate for komatiite crystallization. Bickle's final point related to alteration. He stated that komatiites from Belingwe have only 1-2% H_2O, and that N. Arndt quotes up to 99% fresh olivine in these rocks. Even the smallest olivine microlites are preserved. Since these rocks have comparable compositions to more altered rocks, when normalized to anhydrous totals, he feels that alteration is not a major problem in changing komatiite chemistry.

R. A. Hart then presented some data from modern oceanic basalts that shows substantial MgO influx and SiO_2 leaching during hydrothermal alteration. Data for mafic and ultramafic rocks from Barberton show a similar slope on an MgO vs. SiO_2 diagram, which Hart interprets as clear evidence of Mg-metasomatism in the highly altered (up to 16 wt% H_2O) Barberton rocks. Bickle reiterated that the Belingwe greenstone belt contains very fresh komatiites. Nisbet concurred, stating that some Belingwe komatiites even contain optically fresh

glasses, and that primary olivine accumulation trends such as Na vs. Mg exist: These should be disturbed if there was major alteration.

B. Gorman commented that 24% MgO seems a bit low for komatiite liquids considering the data available for glassy parts of komatiite flows. He also asked Elthon whether komatiites, picrites, and Mg-rich basalts could be related by low-pressure fractionation. Elthon stated that these could be related by olivine fractionation, and that in general komatiites may be related to a large number of basalts with substantially lower MgO. In some cases, however, they can be shown to be unrelated in any simple manner.

M. de Wit raised the question of the volumetric importance of komatiite in Archean terranes, and asked Canadian participants to assess the abundance of komatiite in the Canadian Shield. P. C. Thurston responded, stating that: There are only two known komatiite occurrences in the Wabigoon Subprovince; much less than 1% of the exposed bedrock in the Uchi Subprovince is komatiitic; there are only scattered occurrences in the Sachigo belt; and even the Abitibi belt, which is regarded as being anomalously rich in komatiite, actually contains very little when considered in a regional sense. K. Burke wondered why this mattered; the presence of even a small amount of komatiite in Archean terranes is making a statement about the ancient mantle. Both Hart and de Wit were quick to mention the existence of Phanerozoic komatiites such as the Cretaceous-Tertiary occurrence on Gorgona Island. Bickle commented that the crucial question is the MgO content of primitive liquids in the Archean compared to the Phanerozoic. No one appears uncomfortable with 25% MgO for Archean primitive liquids, and Gorgona has about 22% MgO. De Wit pointed out that such difference in MgO may not be sufficiently large to infer major differences between Archean and Phanerozoic mantle temperatures. Bickle stated that we desperately need more compositional data for komatiitic olivines, particularly from fresh samples. No one disagreed.

E. G. Nisbet agreed that Elthon has identified some major points, but that he has used the wrong type of logic. Nisbet stated that the major hole in their own argument, from which they inferred that highly magnesian (32–33% MgO) primitive komatiite liquids existed, comes not from considerations of olivine-liquid Fe-Mg equilibria, but rather from the possibility of major Mg influx during alteration. Elthon responded that even in the freshest of komatiitic rocks, the degree of olivine accumulation is difficult to discern, unless olivine-liquid matching tests are carried out; phenocryst-liquid matching tests indicate that komatiites with 24% MgO are not liquids.

Chairman K. Burke ended the komatiite bloodbath by moving the discussion to other topics. He commented that the usage by Groves and Batt of rift and platform environments (as quoted in Thurston's talk) in discussing greenstone belts is entirely inconsistent with the ways these terms are used by most geologists. Thurston stated that Groves and Batt's usage of the term "platform" refers to widespread shallow-water sedimentation, with correlatable areas as large as 150–200 km of stromatolitic carbonates. Once again, Burke did not seem convinced.

The session continued with a talk by D. D. Van Reenen et al., who described the metamorphic and structural transition of the low-grade Pietersburg greenstone belt into the granulite facies terrane of the southern Limpopo belt (Kaapvaal craton, South Africa). The Pietersburg belt is at least 3450 Ma, and consists of typical volcano-sedimentary units, unconformably overlain by 2800–2650 Ma sediments. The belt is intruded by 2800 Ma gneisses and 2600 Ma granodiorite plutons. These rock units can be traced 60 km to the north into granulite-facies assemblages; the transition also involves a progressive increase in deformational intensity. Van Reenen suggested that the Pietersburg belt was depressed into the lower crust by tectonic stacking. Fluids responsible for producing the observed retrogression are inferred to have been derived by dehydration of over-ridden lower grade rocks during and after thrusting.

Discussion of this paper was initiated by T. Skulski, who asked if there was any evidence for crustal thickening since the Archean. Van Reenen responded that no post-Archean magmatic thickening has taken place. E. G. Nisbet agreed, and added that the Proterozoic Waterburg sedimentary sequence may have added only about 1 km. P. J. Hudleston was curious about the evidence for increased intensity of deformation in the granulite facies terrane. Van Reenen confirmed that there was abundant evidence for high strain in this area, including isoclinal folding. Original contacts between rock units have all been tectonically disrupted and later invaded by anatectites. Granulitic remnants of greenstone belts exist as xenoliths floating in a sea of tonalitic gneiss. De Wit added that the rocks were much more homogeneously deformed on the high-grade side of the granulite isograd. Hudleston asked if the deformation is consistent with thrusting. De Wit responded positively. Van Reenen added that during convergence and thickening, slices of granulites were juxtaposed against lower grade rocks, but the initial thrusts were overprinted by later granulite grade metamorphism. M. G. Jones asked if retrogressive features were also present in the tonalite gneisses. Van Reenen hesitated, but confirmed that hypersthene was retrogressed to amphibole in these rocks as well. A. H. Wilson remarked that the olivine shown did not resemble spinifex. Van Reenen stated that M. Viljoen and C. Anhaeusser were quite happy to accept these as igneous textures. J. A. Percival asked if there was any evidence for extensional features; Van Reenen said there was none.

The session continued with two contributed talks on thermal considerations. P. Morgan discussed the thermal implications of greenstone belt metamorphism, and presented an alternative thermal model for the Archean continental lithosphere. The common occurrence of supracrustal rocks (including those of greenstone belts) as high-pressure granulites presently underlain by normal thicknesses of continental crust cannot be easily explained without extensive melting during the granulite metamorphic event. Maximum metamorphic temperatures attained for these terranes, therefore, are buffered by the solidus, regardless of age. It is thus impossible for Archean metamorphic assemblages to record higher temperatures than those of modern granulite terranes unless the solidus has changed with time. The necessarily higher global heat loss during the Archean should not then be recorded as mineral assemblages indicating higher geothermal gradients than peak modern gradients, although these conditions may have been more widespread during the Archean. The higher Archean heat loss must also

be reconciled with the existence of diamonds of Archean age, which seemingly require thick (> 150 km) Archean lithosphere. Morgan's solution to this apparent paradox is an Archean continental lithosphere with no asthenospheric heat flux into its base. The thickness of such a lithosphere is independent of global heat loss, so the Archean Earth must have lost its heat elsewhere, for example, through oceanic regions or a different kind of continental lithosphere. The internal radioactive heat production distribution necessary to account for such a situation requires a small but significant amount of heat production in the mantle portion of the lithosphere, and results in a geotherm asymptotic to the asthenospheric adiabat.

Bickle opened the discussion by commenting that Morgan's model of mantle lithosphere enriched in U and Th should leave a recognizable Pb isotopic signature, for example, in Archean mantle xenoliths. He also stated that although the model is conceivable, it should result in high metamorphic gradients in the overlying crust. Morgan disagreed, restating that his model is based on modern-day gradients and modern-day reduced heat flow values. K. Burke expressed uncertainty as to how lithosphere could be stabilized with higher heat generation from below. Morgan responded that heat would not be transferred to the lithosphere from underlying mantle if their temperatures were the same at the boundary. Bickle wondered how fast Morgan's model lithosphere would cool. Morgan responded that his model is sufficiently similar to the present-day situation such that no cooling would be necessary. Burke added that this seems similar to the problem in explaining why old ocean floor stops declining in elevation; McKenzie and Weiss called upon additional convection to keep it high. Bickle returned to the question of crustal gradients, agreeing that lower crustal temperatures were buffered by melting, but he pointed out that wet melting curves, as quoted in Morgan's abstract, may not apply because the lower crust is probably too dry. Buffered temperatures would therefore tend to be greater. Morgan responded that no melting curves were given in his abstract, and that in any case, the choice of melting curves is irrelevant because even the dry solidus would be exceeded in the lower parts of *thickened* crust that produces granulite mid-way through the section. Bickle commented further that if lower crustal gradients are buffered by melting, we should focus on metamorphic conditions of lower grade rocks to examine Archean *conductive* thermal gradients. Although the data is sparse, he cited several examples of greenschist-amphibolite grade terranes (e.g., Pilbara, Yilgarn, Isua) where lower thermal gradients, essentially equivalent to Barrovian conditions, were documented from metamorphic assemblages. Morgan pointed out that even these gradients are likely to be convective rather than conductive. Bickle stated that it depends where they are measured. Burke commented that Andean and Alpine gradients are nowhere conductive. Bickle pointed out that heat flow measurements in the Alps are complicated by perturbations due to near-surface water, but that below this zone the gradient could still be conductive. Morgan added that in places where these problems can be overcome, the resulting gradients are too high to avoid crustal melting during thickening. Areas such as the Rio Grande rift, which can be successfully modeled in terms of the underlying Socorro magma chamber, indicate that heat flow is not conductive. Bickle agreed that many

metamorphic terranes exist where magmatic heat transport is very important, and this is easily documented by elevated temperatures at low pressures, but there are other terranes where convective heating is not important. Burke requested that Bickle name five such terranes. Bickle offered two (the Barrovian sequence of the Scottish Dalradian and the European Alps, except for some of the thermal domes) before being cut off by Burke, who exclaimed that once thermal domes are recognized, gradients are convective, not conductive. Bickle stated that the convective parts of these terranes (the thermal domes) can be recognized and separated from the conductive, low-temperature, high-pressure parts. These can be used to make estimates of the heat influx at the bottoms of those crustal sections. Such perturbed (convective) parts can be recognized in Archean terranes as well as in younger ones. Burke was not convinced that there were substantial parts of the world where metamorphic assemblages did not involve magmatic heat transfer.

D. Abbott asked Morgan if his model could be tested by looking at hot-spot thinning of the continental lithosphere. Morgan responded that since cratons do not generally show that phenomenon, they are resistant to hot-spot thinning. He indicated that not *all* lithosphere should be expected to have been similar to that described in his model; only if the proper conditions were achieved would such lithosphere become stabilized. Most continental lithosphere is probably different, and might be subject to reactivation. Burke asked if what Morgan was inferring was that everything that has not been reactivated could not be reactivated. Morgan responded that there must be a reason for selective reactivation. Burke stated that all reactivation is related to collisional or Andean margin type activity.

Nisbet commented that Morgan's model has an inherent contradiction: The lithosphere is enriched in heat-producing elements, yet it must also be refractory to keep it stable at those high temperatures. The model is therefore transient because eventually the heat-producing elements will tend to escape upward into the crust. Morgan admitted that his model is not totally stable and that the distribution of lithospheric heat production will change with time. Burke interjected that such situations may be transient, but only on a 5-Ga scale; although the Kaapvaal craton was assembled 3.4 Ga ago, not much has happened to it since.

Wyllie reminded us that lateral heterogeneities in the mantle may have relevance to the question of komatiite genesis. These might be produced by high degrees of melting at local hotter zones associated with upwelling parts of convection cells. Thus Archean simatic melts may have been derived from two sources: those that gave rise to komatiites, and those that produced near-contemporaneous basalts in "standard" fashion. Burke endorsed Wyllie's comments as being useful in reminding participants not to consider the Earth as being (or having been) radially symmetrical.

D. Abbott and S. Hoffman discussed the possible importance of ridge subduction and hot spot abundance in greenstone belt formation. They feel that the one consequence of greater internal heat production of the Archean Earth may have been a greater proportion of subduction of young oceanic lithosphere, and that therefore ridge subduction would have been more common.

They speculate that Archean ridge subduction may have been dominated by the type in which oceanic lithosphere comprised both the overriding and subducting plates, a present-day example of which lies in the Woodlark basin of the western Pacific ocean. Features of this system possibly applicable to Archean greenstone belts include basalts with arc affinities and closely spaced volcanic edifices in the associated island arc, which could thicken layer 2, leading to higher probability of hydrothermal alteration and metallogenesis. Another consequence of a hotter Archean Earth may be a greater abundance of hot spot activity, which may have increased the incidence of buoyant subduction. Abbott outlined in detail the importance to their model of the flexural bulge of the subducting plate.

In discussion, J. F. Casey pointed out that there are other examples of oceanic plates with ridge subduction, such as the Aleutians. Abbott agreed and stated that high-Mg andesites occur on the continental plate. Other examples include the Kula ridge, the Gulf of California, and southern Chile. She doubts, however, that Mg-andesites would be found in ophiolites formed at mid-ocean ridges. Casey mentioned that high-Mg andesites occur in forearcs such as the Marianas and Japan. He wondered further how Abbott's model resulted in ophiolitic obduction in an environment of ridge subduction. Burke suggested they may be analogous to slices in an accretionary wedge. Abbott agreed, pointing out Eldridge Moore's proposal that the Troodos and Samail ophiolites represent obducted pieces of hot, young oceanic plates, essentially equivalent to ridge crests. De Wit asked how Abbott accounts for the almost total absence of high-Mg andesites or boninites from greenstone belts. He could think of only one place from which they may have been reported. Abbott commented that K. J. Schulz reported high-Mg andesites from northern Minnesota, where they occur late in the sequence, and although not very abundant, they are quite similar to some reported in Japan. Elthon expressed his surprise about Abbott's reluctance to accept the high Na and Ti contents of Woodlark basin rocks as being magmatic. Abbott informed us that such rocks are dredged only from fracture zones, where processes additional to those occurring at ridge crests might be taking place. She could not account for these rocks in ways other than their being related to spilites. Elthon suggested that they may have formed by remelting hydrothermally altered rocks in fracture zones, but he does not doubt that their compositions are magmatic. Abbott accepted this. Schwerdtner inquired if Abbott's flexural bulge model was Newtonian and whether it was linearly elastic. Abbott confirmed this, and added that if the model was made more viscoelastic the numbers for the stress effect would probably decrease, but it would also increase the thickness of the layer, so it is difficult to say which effect would dominate. Discussion on this question continued, but it was inaudible on the videotape. Abbott reiterated that there must have been, on average, more young ocean lithosphere subducted in the Archean. Bickle commented that there may have been more melting to deeper levels in the Archean: this could have led to faster spreading. Both Morgan and Abbott, however, have suggested hot spot heat loss may have been an alternative mechanism. Burke commented that if thickened oceanic crust such as that in oceanic plateaus were obducted, then the sliced-off fragments might contain only basalts and not ultramafics. This has been documented in the Caribbean, in southern Malaita, and in Alaska.

Thurston changed the topic of discussion by commenting that the occurrence of ash-flow magmatism from 3.0 Ga onward implies compositionally zoned magma chambers and, in turn, a relatively constant rate of basaltic magma supply to those chambers. If basaltic magma were supplied at rates higher than about 100 km^2/yr, then magma chambers would become dominantly mafic, and incapable of generating ash-flow type volcanism. Burke commented that this is a statement about a particular environment, and that others occur in greenstone belts. Burke wondered further how these considerations compare with the present-day situation, and Thurston replied that volume-periodicity relationships indicate much the same conditions in the Archean.

De Wit then changed the discussions to cratonic areas, particularly about the transitions from greenschist to granulite facies. He asked Van Reenen if there is any change in fluid inclusions from the hydrated to the dehydrated terrane in his area, as has been documented in India and Kapuskasing, Ontario. If there is evidence for abundant CO_2, where does it come from, and what are the implications for apparently associated Au mineralization? Van Reenen responded that his high-grade rocks are similar to those from India, described by Newton and Hansen. They are completely dominated by CO_2, with less than 0.2% H_2O, and there is a progressive change in density of pseudosecondary inclusions from south to north across the orthopyroxene isograd. Van Reenen stated that the hydration of orthopyroxene took place at temperatures as low as 600°C. In his view, the carbonic fluids were derived from decarbonation of typical greenstone belt lithologies during and subsequent to overthrusting. Associated shear zones are dated between 2450 and 2600 Ma, and are accompanied by CO_2 metasomatism. Gold occurrences seem to correlate with these shear zones rather than with lithology. De Wit asked if Van Reenen believed the CO_2 to be indigenous to the crust, rather than mantle-derived, and Van Reenen said yes. Wyllie asked about the abundance of limestone in greenstone belts, and Van Reenen said there was very little, but that calc-silicates are abundant. Wyllie then wondered where the original CO_2 came from; Van Reenen offered the possibility that it was derived from hydrothermally altered greenstone belt lithologies. De Wit concurred about the abundance of metasomatically altered material (carbonated ultramafic), and added that the CO_2 may ultimately have been derived from the mantle. Burke asked for specification as to how much carbonated rock existed in these belts; de Wit estimated between 1–10%. Burke felt that in this case there would be enough CO_2 available if the belts were thickened. C. Schiffries added that calculations by J. W. Valley indicate that only 1–2% carbonate would be needed in low-grade lithologies to account for the CO_2 influx during granulite metamorphism, and that carbon isotopes show evidence for crustal reduced carbon. The morning session was then adjourned by Chairman K. Burke.

J. A. Percival and K. D. Card started off the afternoon's **Session III** (Greenstone Belts: Their Boundaries, their Surrounding Rock Terranes, and their Interrelationships) with an invited keynote paper. They reminded the workshop participants that the major controversies about the tectonic

environment in which greenstone belts were formed exist because good evidence for specific settings, such as within continents or oceans, is rare. They believe, however, that the existing data shows Archean volcanic sequences to have much in common with Cenozoic volcanic arcs. Accordingly, Percival and Card expressed their belief that most greenstone belts, and specifically those in the Slave and Superior Provinces of the Canadian shield, were formed along Andean or Pacific-like convergent margins. They also noted that a significant alkaline component of the igneous rocks suites of the Superior Province may be related to collisional events. With respect to the original relationships between greenstone belts and their surrounding terranes (which they identify as either metasedimentary belts or granitoid terranes), every conceivable type of structural, igneous, and stratigraphical contact appears to have been recorded, but they noted in particular a major difference between terrane relationships in the Slave Province and the Superior Province. In the former, stratigraphic onlap exists between the metasedimentary belt and the metavolcanic (greenstone) rocks; in the latter, the metasedimentary belts alternate with the volcanic-plutonic belts along the tectonic contacts. Percival and Card briefly reviewed both the plutonic terranes and the metasedimentary belts. They noted that in the plutonic terranes, many tonalite-diorite plutons are coeval with the volcanic hosts, as determined by dating of abraded zircons ("precise zircon dating"). Also, plutons external to the greenstone belts are generally similar in composition and age to plutons within the belts. Although some of the plutonic rocks are older (and represent basement to the greenstone sequences), contacts are generally intrusive or tectonic. Plutons of granodiorite-granite compositions commonly post-date the youngest volcanic rocks of the greenstone belts and their tectonism by 5–25 Ma. Percival and Card believe that the tonalitic magmatism may be genetically related to shallow angle subduction, while the equally voluminous granodiorite-granite magmatism may be due to silicic melting during crustal thickening as a result of collisional or accretionary events. They emphasize, however, that collisional processes between Precambrian blocks have not yet been substantiated by paleomagnetic studies.

The metasedimentary belts referred to by Percival and Card consist predominantly of turbiditic graywacke and shales. They constitute a significant component of the Slave and Superior Provinces. Those of the Slave Province have been interpreted as part of accretionary prisms; these in the Superior Province as narrow, elongate transtensional basins along the major tectonic breaks between crustal blocks. Finally, Percival and Card expose and comment on what is presently known about the transitions between low-grade (greenstone belt terranes) and high-grade (gneiss terranes). They site two types of transitional relationships: (1) where (low-grade) greenstone belts can be traced *laterally* into high-grade metasedimentary belts (e.g., the Wabigoon and Wawa belts can be traced into the Quetico high-temperature/low-pressure terrane); (2) where (low-grade) greenstone belts can be traced *vertically or down-section* into their high-grade amphibolite-granulite (high-pressure/high-temperature) equivalents. Examples of this second type of transition has been clearly documented in the Kapuskasing structure (Ontario), in the Pikwitonei region

(Manitoba), in the northern and southern marginal zones of the Limpopo mobile belt (as traced from their respective low-grade granite-greenstone terranes of the Zimbabwean and Kaapvaal cratons), and in southern India. In the Kapuskasing structure the lowermost exposed rocks were subjected to P-T conditions between 7–8 kbars and 700°–800°C; in the Pikwitonei structure pressures vary between 7–12 kbars (not unlike those from Southern Africa). In the Superior Province these adjacent high- and low-grade terranes are thought to represent different tectonic environments exposed at different crustal levels. The deeply eroded Southern Andean batholith (interpreted as the roots of an calc-alkaline volcanic-arc) flanking the Cenozoic Rocas Verdes back-arc basin is thought to be a Mesozoic paired analogue. This interpretation is different from the proposed models for the high-grade/low-grade transition of the northern Kaapvaal craton: Here differential uplift following continental collision tectonics is believed to be a "best fit" model.

During the following three talks, different examples of relationships between greenstone belts and their surrounding granitoid terranes were examined: details of an igneous, a structural, and an unconformable (stratigraphic) relationship were described in that order; all have profound tectonic implications.

In the first talk, F. Schwerdtner presented his detailed observations on selected granitoid-gneiss complexes from within the Wabigoon and Wawa greenstone belts. Schwerdtner explained how the field data was incompatable with his previous models in which he envisaged the gneiss domes as purely diapiric structures. The new observations clearly indicate that the gneiss domes are lithologically composite and contain large sheath-like structures that are deformed early plutons, distorted earlier gneiss-domes, or early ductile nappes produced by folding of planar plutonic septa. Thus, Schwerdtner concluded that prominant gneiss domes are composed of *prestrained* tonalite-granodiorite and represent the dense hoods of magmatic granitoid diapirs (commonly a syenite-diorite crystal mush). The work also implied that the synclinal-like structures of greenstone belts predate the doming of the granitoids: evidently the early deformational history of the greenstone and granitoid-gneiss domes is far more complicated than is presently understood.

In the second talk, R. L. Bauer *et al.* described similarities and contrasts in structural history and style between the greenstones of the Vermilion district (northeast Minnesota) and the Vermilion granitic complex. The two terranes are separated by faults. Structural analysis in the boundary regions between the two terranes indicates that both sustained an early (D_1) recumbent folding of regional scale. In the greenstone belt this folding is attributed to deformation of soft or poorly lithified sediments. They were able to reach this conclusion because finite strain analysis on clasts in sedimentary units can be completely accounted for in terms of a second deformation (D_2). D_2 formed upright folds and dextral shear-zones during transpression across the belts. Although in detail the structural history of the two terranes is similar, differences in structural styles and a late stage (D_3) structural history are prominent. Bauer *et al.* attribute this to a combined difference in juxtaposed crustal levels of the two terranes during late deformation and increased heat flow in the greenstone belt as a result of late-D_2 plutonism.

In the last presentation, M. Wilks and E. Nisbet described details of a profound (and probably one of the world's best preserved) Archean unconformities between the rocks of a greenstone belt and its surrounding granitoid terrane. The unconformity occurs at the contact between the Wabigoon greenstone belt and a ca. 3.0 Ga tonalite complex. At three localities the tonalite can be traced through a weathered profile into the overlying sediments of the Steep Rock Group. The latter comprises a basal conglomerate, a thick stromatolite-bearing carbonate member, and a manganese-iron-rich ore zone that locally contains aluminous kaolinite and gibbsite (interpreted as a ferruginous bauxite, and indicative of an oxidizing atmosphere); the section terminates with a sequence of ultramafic pyroclastic rocks (22% MgO) that contain thin spinifex-bearing komatiitic basalts (15% MgO). If the Steep Rock unconformity once was more widespread, it has apparently been obliterated elsewhere in the Wabigoon greenstone belt during intense tectonism.

P. Hoffman kicked off the discussion by asking F. Schwerdtner if it was possible to establish the kinematics of shear associated with the development of the lineations in the rocks he had described. Hoffman pointed out that he would be most convinced about the diapirism if Schwerdtner could demonstrate an "upside down" sense of shear on opposing sides of the dome. Schwerdtner didn't know if he could approach the kinematics using the lineations, but he tried using the folds. He reiterated that the leucocratic center of the gneiss dome was fringed by a mixed zone containing relics of greenstone lithologies; in this envelope there is abundant folding. Schwerdtner said that he had studied the asymmetry of several hundreds of these folds, and although fold styles are variable and there is abundant refolding, the results indicated that the dome had risen relative to the surrounding areas. The test was only completed around two-thirds of the dome's periphery for lack of enough folds along parts of the dome. Hudleston commented that the lineation in Schwerdtner's dome was the result of cumulative deformation, much of which appeared to have predated the doming. Schwerdtner agreed, but pointed out that he could not be sure how much of the strain was earlier than doming; he had not encountered the type of strain markers needed (i.e., features of known geometry at the start of doming) to quantify this accurately. He explained that the structural map represents only the total strain, but he wasn't at all convinced yet whether or not the lineation entirely predated the doming. De Paor asked if it is necessary to involve two phases of deformation; he commented on the fact that the strain pattern looked as if it could have developed in one phase if an anticyclonic vorticity was applied to the incremental strain field. Schwerdtner replied that this would only apply if the domal structure was prefectly symmetrical, which it was not, and he had therefore not considered this possibility. W. R. Muehlberger had noted that all the folds in the region of the dome area had an "S"-shaped geometry. He wondered if the doming had not occurred during regional simple-shear. Schwerdtner thought that this was unlikely since the shearing was dextral in the area just to the south (near the Quetico fault), and one would expect "Z" folds. De Paor remarked, however, that it was not uncommon to find areas of dextral and sinistral shear zones (or wrench faults) in close proximity of one another.

Schwerdtner had no objections to having regional shear during doming. Again, his main point was that if one started with an *undeformed* rock and deformed it into subcircular structure, the structural patterns that he had documented could not be accounted for. An earlier structure must have been there before the rising of the dome: It may have been a synchronous process, but an earlier strain *had* to have been there before the subcircular structure development. Hudleston tried again— "Could it not have been a synchronous process as D. De Paor suggested?" Schwerdtner was adamant and could not see how that was possible and he doubted if there was any experimental way of showing it. De Wit asked if Schwerdtner might not consider more complicated modeling using a finite element difference approach that incorporated questions posed by De Paor and Hudleston. It was certainly true that during the course of this meeting we have learned that there was a lot of horizontal shortening across greenstone belts. It would be interesting, for example, to compare field patterns with model patterns of a system that integrated diapirism with overthrusting. Schwerdtner agreed. D. Abbott asked Schwerdtner if he could comment on the rheology necessary to develop domes; Does it require a certain set of viscosity contrasts? Would juxtaposed rocks of the "wrong" contrasts not sustain any doming? Schwerdtner replied that given the presence of dilatant openings, diapirism would develop even if the (lowermost) medium was dense (provided it was very fluid). In these cases the fluid would not rise to the surface; it would only rise until it reached equilibrium between the overburden pressure and the weight of the diapiric column. Schwerdtner believes that there are examples in some areas where this might have happened: magmatic crystal mushes, for example, rising into fault zones and other dilatant structes. Such fluids are not Newtonian; they are either Bingham bodies or behave as nonlinear (sometimes called pseudo-plastic) fluids: they all have a yield point, however low. Martyn then asked if you could not produce the lineations by crustal extension and at the same time produce a (sedimentary) cover into which the diapirs (of the stretched basement) could rise isostatically, again a continuous process. Schwerdtner answered no, because in the area there is a vertical lineation that formed by transverse compression; he went on once again to emphasize that he could not properly determine the path of deformation of the body itself from the finite strain patterns. We need more clues to establish the increments of deformation that collectively led to the final results. He said that he was working on this, using conventional structural analysis of boudins, veins, folds, porphyroblasts, etc. He outlined one new interesting approach that he was investigating: the use of intensity of rock anisotropy, such as lineation or foliation. He explained that, for example, if a rock develops a gneissic fabric to such an intensity that it can no longer contract (redeform) uniformly along its strain trajectories, any following stress increment that tries to compress that lineation will buckle it. Schwerdtner believes that is a type of increment that you can document and work with to decipher the deformation path.

K. Burke pointed out a more general problem related to granitic intrusions and possible diapirism in mountain belts: D. Hodge (S.U.N.Y. Buffalo) had been examining the problem of partial melting of a continent and separating a granitic minimum melt and had found that the viscosity contrasts were so small

between a minimum melt and warm continental rocks that it was hard to get the melt material to move up; a crustal fracture had to be created in order to get the granitic fractions to rise. The comment was then transformed into a question. Did Schwerdtner have any idea of the original shape of the granitic material before it became a diapir? Whereas in salt-dome terranes it was well known that the original salt was horizontally bedded, such an inferred geometry could not easily be inferred granitic environments. Schwerdtner replied that he did not know.

P. Hoffman commented that the succession described by Wilks and Nisbet was apparently very like those of the early Proterozoic platform sequences. He wondered if the late stage subsidence recorded in the Steep Rock sequence might be explained by flextural loading, perhaps by tectonic thickening of a thrust belt. If so, the uplift needed to form the bauxites in this succession might be due to a migrating flextural bulge. Hoffman stated that in the absence of evidence of faulting, such a model might be more realistic than one that involved subsidence due to crustal extension. M. Wilks replied that there is normal faulting in the area. Hoffman asked if these faults perhaps post-date the stromatolite formation and pre-date the iron formation. Wilks replied that normal faults offset the entire Steep Rock succession, and that, interestingly, the mafic dikes intruded along the fault planes. These dikes are thought to be feeders to the Ashfall Formation, so the faults must have been present prior to their deposition.

Hoffman asked Percival if he could distinguish between an inter-arc rift to account for the metasedimentary belts as opposed to unrelated arc collisions that had caught up accretionary prisms or peri-arc basins. Percival replied that the only evidence he could cite to substantiate a rift-type of environment was the high heat flow implied from the metamorphic assemblages. He stated that accretionary prisms tend initially to have very low heat flow (low-T, high-P blueschist facies metamorphism), and were often later overprinted by high-temperature/low-pressure conditions. It was not clear if such evolved metamorphic patterns were present in the metasedimentary belts. D. Davis remarked that he found it interesting that people reach opposing conclusions using the same data base. For example, Garth Edwards used Davis' U-Pb zircon date to try to test the idea that the Superior Province represents a progressive accretion of island arcs that grow younger from north to south across this terrane. However, if you look at Davis' poster session, which portrays at least 200 analyses, you will see that between the period of 2750 Ma and 2700 Ma there is simultaneous volcanic and plutonic activity over the entire Superior Province. He believes that this totally discredits the model of progressive accretion. Moreover, there are other detailed studies (such as in the Wabigoon greenstone belt, with more than 60 analyses) that imply that there is absolutely no evidence of older material in these rocks. This strongly suggests that they developed in ensimatic environments, and it could be that they developed as an island arc system. On the other hand, there are marginal conglomerates in the greenstone belts that flank gneiss belts (i.e., the English River belt and in the central part of the Wabigoon subprovince, both of which are ca. 3.0 Ga), which suggest that the greenstone belts developed adjacent to continental crust also. Davis stated

that he therefore favored a model of widespread cratonic rifting throughout the Superior Province; he saw no role for subduction at this point. M. Bickle asked if there was agreement that the younger stages get younger to the south. Davis replied that there was only a secular variation in that the youngest volcanic sequences (2710–2700 Ma) occurred in the southern belts (i.e., the Wawa and the Abitibi belts). A general discussion followed as to whether or not the geochronological patterns could be interpreted as a result of colliding arcs. Davis did not think so, but there was no overall agreement among many of the Canadian participants.

This session continued with the final four contributed talks. D. R. Hunter et al. presented the results of their work on the southeastern section of the Kaapvaal craton. The talk was given by A. H. Wilson. In southwestern Swaziland, a > 3.5 Ga suite of tonalite-trondhjemite gneisses and amphibolites comprises the oldest-dated sialic rocks in the Kaapvaal craton, and may represent the basement on which younger greenstone belts accumulated. Evidence includes a complex structural prehistory not seen in the greenstone lithologies, and structural superposition of greenstone on gneisses. The abundance of metaquartzites and metapelites in the greenstone sequences supports a nearby sialic basement. Greenstone remnants include different proportions of komatiite, high-Mg basalt, tholeiite, subvolcanic intrusions, clastic and chemical sediments, rhyolitic air-fall tuffs, and flows. The differences between these remnants represent either different exposural levels and/or ages of accumulation. Isotopic data at present cannot distinguish between formation at ~3.6 Ga with resetting at ~3.1 Ga, or formation at ~3.1 Ga with contamination by 3.5 Ga sialic crust. The terrane was intruded by mantle-derived tonalites and an anorthositic layered intrusion (3.3 Ga?), by sheet-like granitoid batholiths (3.2-3.0 Ga), and a potassic granite batholith (3.0 Ga). After this period, uplift, weathering, and minor volcanism took place, coinciding with the beginning stages of the thick rift-like sediments that constitute the Pongola intracontinental rift (~3.0-2.8 Ga).

R. E. P. Fripp then presented a tectonic synthesis of the Yilgarn Craton of western Australia, which contains about 70% granitoid and 30% greenstone. Granitoids include pre-, syn-, and post-tectonic types. The youngest of these (~2.6 Ga) are temporally equivalent to the youngest greenstone belts, which are up to about 2.8 Ga. Although most greenstone-granitoid contacts are tectonic, most workers agree that pre-greenstone sialic basement is preserved in places. The greenstones have been interpreted in terms of three large basinal structures, one of which is considered to be a rift, but Fripp presented structural data that show that when restored the entire greenstone package might have constituted a single basin. There is structural evidence for fold-nappe and thrust-nappe tectonics as well as large-scale imbrication or slicing, not unlike that seen in young fold belts.

G. N. Hanson, E. J. Krogstad, V. Rajamani, and S. Balakrishnan discussed the Kolar schist belt and surrounding gneiss terranes of south India. The talk was given by Krogstad. The schist belt itself evidently represents a discontinuity between two granodioritic gneiss terranes with different ages, structural styles, and compositions. East of the schist belt are relatively homogeneous granodiorite gneisses intruded at 2529 ± 1 Ma,

and metamorphosed to amphibolite grade at 2520 Ma. To the west of the belt are older (2610–2550 Ma), more complexly deformed gneisses, some of which show evidence for contamination with 3200 Ma basement. In the schist belt, two suites of komatiites and tholeiites can be distinguished on the basis of REE, but Sm-Nd isotopic data from both suites lie along an "isochron" of about 2900 Ma. Hanson and colleagues interpret the Kolar schist belt as marking the site of a *suture* between two late Archean continental terranes.

S. M. Naqvi ended the session with a talk on the tectonic evolution of the Chitradurga schist belt of south India. About 80% of this belt consists of detrital and chemical sediments, including conglomerates, quartzites, graywackes, shales, phyllites, carbonates, banded iron formations, and banded manganese formations. The graywackes represent debris derived from gneissic and K-rich granitic sources. Rare earth element patterns of these rocks have both positive and negative Eu anomalies, an unusual feature among Archean sedimentary rocks. The volcanic rocks of the belt include ultramafic komatiites as well as mafic, intermediate, and acid volcanics. Most of the belt is greenschist facies, but locally reaches amphibolite and granulite grade. Structures indicate that horizontal compression, possibly related to collision tectonics, played a major role in the development of the belt. Naqvi offered two possible tectonic models for the Chitradurga belt: (1) development in rifted Archean continental crust, followed by collapse, shallow subduction, and horizontal compression; or (2) development on oceanic crust, followed by horizontal motion and welding of two separate continental blocks along a suture.

The discussion was opened by J. Vearncombe, who asked Wilson to clarify what he meant by "stabilization" of crust. He suggested that care must be taken not to give the impression that the Kaapvaal craton was "stable" since 3.0 Ga. He pointed out that there are five major Precambrian sedimentary basins, major thrust events in the Witwatersrand basin, intrusions like the Bushveld, and major Proterozoic transcurrent faulting with substantial associated mylonites. Wilson responded that he was only referring to events in his area (i.e., the southern part of the Kaapvaal craton); his results should not be taken as a general model for the entire Kaapvaal craton. K. J. Schulz asked Wilson about his evidence that the oldest event in his area pre-dated the Barberton belt. Wilson responded that the isotopic data is at present equivocal. Although no basement to the greenstone belts is presently observed, they assume a pre-3.5 Ga sialic crust existed, based on Hunter's work in the ancient gneiss complex of Swaziland. Naqvi asked Wilson to clarify what he meant by "initial" vs. "final" stabilization. Wilson explained that final stabilization referred to a large influx of granitic material that stabilized the crust on which the Pongola sediments were deposited. Initial stabilization refers to a crustal thickening event during which earlier sialic material was deformed. Hoffman inquired if the ages Wilson quoted were from zircons or Rb-Sr isochrons. Wilson said that some were Rb-Sr, others Pb-Pb, and he reiterated that at the present time the geochronology is equivocal. Two good Pb-Pb isochrons on komatiites from Nondweni give ages of about 3.15 Ga, whereas Sm-Nd data on the same rocks gives an age of about 3.5 Ga.

Nisbet asked Naqvi to elaborate on the REE signature of the sedimentary rocks he described in terms of Taylor and McLennan's crustal evolution model. Naqvi explained that although Taylor and McLennan's work indicates that graywackes older than 2.8 Ga do not have Eu anomalies, graywackes and associated shales of *all* ages (e.g., 3.5–2.5 Ga) from Chitradurga have negative Eu anomalies. These anomalies appear to reflect incorporation of secondary minerals at the time of deposition, rather than a signature from their source areas. Ludden inquired about the evidence for this. Naqvi explained that Eu anomalies in these rocks can be correlated with other chemical features: shales or graywackes with high Fe content have positive Eu anomalies, whereas those with high K_2O have negative Eu anomalies. He illustrated this with a slide showing spectacular negative Eu anomalies in graywackes older than 2.9 Ga. Ludden asked about the REE patterns of older tonalites and trondhjemites from this area; Naqvi responded that those rocks have smooth patterns with no Eu anomalies.

De Wit asked Fripp if he thought that, on a large scale, the entire Yilgarn craton represents a section through a fold and thrust belt, and if so, what was the polarity of the thrusting. Fripp responded positively to the first part of the question, but said that kinematic data are not available yet to determine thrusting directions.

T. Barrie asked Krogstad if the volcanic rocks used to construct his Sm-Nd isochron were cogenetic. Krogstad explained that the komatiitic amphibolites from the western Kolar can be modelled using trace elements as having been derived from a similar source with varying LREE depletion. The eastern amphibolites are not necessarily associated; they appear to have been derived from an enriched source. Since the eastern amphibolites have a galena model age of about 2900 Ma, there is some evidence to support these as temporally equivalent to those from the western side of the belt. A more satisfactory method is needed, however, to determine ages of amphibolites. Ludden pointed out that recent data from Kambalda show the danger of interpreting Sm-Nd isotopic data in terms of isochron relationships; there they have been demonstrated to be mixing lines. Krogstad agreed that care must be taken to show comagmatism between rock samples before extracting age information from Sm-Nd data. The session was adjourned by Chairman P. Hoffman.

V. Greenstone Belt Tectonics: Synthesis and Destiny

Introduction

During the evening following the close of the technical session, the participants were divided into the working groups that met to evaluate critically the ideas and data presented at the workshop, formulate or reformulate a set of important questions, and outline what should be done in future research efforts to answer some of these questions. Working Group I consisted of 18 student participants; Working Group II included 38 of their ostensible mentors. The conveners felt that it would be fruitful for the students, who represent the future as well as the present of greenstone belt research, to meet separately, and thereby become better acquainted with one another, and have an opportunity to discuss those matters in an environment free of influence from their teachers. This turned out to be a good idea. The conveners, perhaps injudiciously, decided not to attend either of these meetings. Each of the groups was asked to select a spokesperson, who would then summarize the delibrations in a short presentation in the next morning's plenary session. T. Skulski was selected to represent Group I, and R. E. P. Fripp presented Group II. Written summaries of these group discussions are given below.

Working Group I: Recommendations and Suggestions

1. The time has come for adopting a quantitative approach to solving Archean geological problems. This requires the use of precise and accurate terminology in describing what is observed.

2. There is a need for objective mapping of greenstone belts.

3. We recognize the need for the development of chronos-tratigraphic columns in Archean greenstone belts. Archean marker beds are more difficult to delineate given the absence of fossils and poor state of preservation. Time lines should be developed utilizing geochronology (high-precision U-Pb zircon techniques) and marker beds such as air-fall units, ignimbrites, and plagioclase-megacryst bearing lava flows.

4. We have to quantify the depth of depositional basins. Can we say anything better than above or below wave base? Pillow vesicles in glassy rinds may have applications here.

5. We need to quantify the true and original thickness of Archean supracrustal successions. Strain and kinematic analyses can help constrain this.

6. Potential method geophysical techniques can complement field mapping in constraining the configuration of greenstone belts.

7. There is a consensus that the understanding of greenstone belts cannot be achieved independent of their adjacent high-grade terranes.

8. What is the significance of hydrothermal alteration in Archean greenstone belts? Detailed studies of associated sulphide mineral deposits may enable inferences to be made on tectonic setting.

9. A critical evaluation of the role of komatiites in greenstone belts is required because of its implications on crustal evolution and the Earth's thermal history. The following problems have to be addressed:
   extrusive vs. intrusive relations (Barberton?)
   role of contamination
   origin of spinifex texture (Barberton?)
   extent of Mg-metasomatism

10. We need to know what the source regions are for Archean sediments. Detailed studies of clastic materials can provide information on source areas, some of which may no longer be present. When integrated with facies analyses these studies provide constraints on granite-greenstone evolution.

11. The validity of uniformitarianism can be tested by comparing Phanerozoic tectonic environments with Proterozoic mobile belts and Archean granite greenstone terranes.

Working Group II: Recommendations and Suggestions

Working Group II concurred that more detailed field work was needed to build on our incomplete understanding of Archean processes. Many of the secular changes that have been proposed, for example, the geochemical signatures of sediments and the relative abundances of "Al-depleted" lavas and K-rich granites, may not be as convincing as first thought. Most workers now seem to agree that modern tectonic processes are appropriate analogs for the Archean. It must be emphasized that greenstone belts are not just Archean! The group made the following recommendations for future work:

1. Compare Archean greenstone belts with possible modern analogs, including: southwestern Japan, Coast Ranges (British Columbia), Sierre Nevada (Klamaths), Central Newfoundland, Eastern Australia (Paleozoic), Rocas Verdes (Chile), active arcs and rifts such as the U.S. Basin and Range, and the Taupo Volanic Zone (New Zealand).

2. Future conferences could be held on structural techniques. Field workshops emphasizing modern tectonics (e.g., ophiolites, rifts, arcs) were a popular recommendation.

3. Future research should involve an integrated approach, including high-resolution geochronology, structural/stratigraphic work, geophysical techniques, remote sensing, metamorphic petrology, and theoretical modeling.

Spokesmen Skulski and Fripp gave short presentations summarizing these recommendations, and short disciplinary summaries were then given by representatives selected by the conveners. These summarizers were asked to submit short written version of their viewpoints, which are included below.

There was extensive discussion among nearly all of the participants during this final session. We have not, however, included a detailed account of these discussions because most of the points raised can be found incorporated in the written contributions that follow.

A Tectonic Viewpoint

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Many different tectonic environments were described from greenstone belts during the meeting and all aspects of the Wilson cycle of ocean opening and closing appear to be represented. A breakthrough was the description by Wilks and Nisbet of what appears to be a fragment of a well-developed Atlantic-type or rifted continental margin from Steep Rock Lake (Ontario). Sandstones and conglomerates there overlie a tonalitic basement and are themselves overlain by up to 500 m of stromatolitic carbonates above which a laterite bears witness to the oxidizing character of the terrestrial atmosphere about 2.8 Ga ago. Steep Rock Lake lies on the southern margin of the highly tectonized Wabigoon belt and appears to be an isolated, well-preserved fragment of an Archean rifted margin that has elsewhere been completely obscured by later deformation.

The idea of overwhelmingly intense deformation (which only a few small areas escape) as the dominant feature of greenstone belts emerged from many of the talks and especially from the poster presentations of detailed field mapping. Numerous participants referred to rift environments, although as Burke and Sengor pointed out, these would have to be perceived (in the intensely deformed greenstone belts) through obscuration by later processes. Some contributors spoke of "ensimatic rifts" or of "narrow ocean basins" (e.g., Davis *et al.*), but all oceans are narrow twice: once when they first open and once just before they close. There is no way of telling from a suture zone how wide the ocean was that has closed to form the suture. De Wit presented superb detailed maps of an ophiolite approximately 3.6 Ga old from which it would appear that, however wide or narrow the Archean oceans might have been, the rocks underlying them were very like those of today's ocean.

Thurston and Ayres described the bimodal volcanic rocks and related volcanic structures of the Superior Province, which closely resemble those in the rifted crests of Andean volcanic arcs (e.g., the Taupo province of New Zealand). An alternative analogy that they also suggested (to the Rio Grande rift) seemed less likely. It is probably significant that Andean margins are prominently represented in greenstone belts because of the huge volumes of igneous material (representing both that newly added to the continents and older reprocessed crust) in these areas.

Features formerly thought to be distinctive features of greenstone belts (e.g., confinement to the Archean, occurrence of very thick continuous sections, and occurrence of sediments without europium anomalies) were shown to be unreal by numerous authors and the old question "What was distinctive about the Archean?" was answered "much less than some of us thought ten years ago."

The thermal state of the Earth was clearly different. Heat was generated at a much higher rate in the Archean than it is now. Komatiites are probably evidence of this, but some participants (e.g., Elthon and Hart) indicated that they might represent less extreme conditions than had been widely thought.

Although the workshop profited from full discussion and frank exchange, there is an obvious need for more of this kind of dialogue betweeen structural geologists, petrologists, isotopic workers, and theoreticians. Discussions in the field are likely to prove particularly stimulating, especially in the best-exposed areas of Africa, India, and Australia.

A Geochronological Viewpoint

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The Archean presents a number of difficulties for geochronology. It is now well known that the application of "soft" geochronological techniques such as K-Ar and Rb-Sr mineral isochrons do not give ages for igneous emplacement but record some later metamorphic overprint. At an early stage it was hoped that this problem could be overcome by the use of Rb-Sr whole-rock isochrons or U-Pb dating of zircons, which were thought to be able to penetrate the "veil of metamorphism" and record primary igneous events. The precision of the whole-rock method is generally limited to a few percent of the age because of the limited spread in Rb-Sr ratios that can be generated within a suite of rocks. This is considered to be sufficient, however, for resolving orogenic events. Uranium-lead analysis of zircon offered the potential for more precise ages, because of the use of more than one decay system and the fact that the daughter element is almost entirely radiogenic. Zircon is, however, subject to recent, partial lead loss, indirectly due to the accumulation of radiation damage that renders the crystal chemically reactive and subject to later alteration (Krogh and Davis, 1974). Because of the multiple decay systems lead loss can be corrected at the expense of reduced accuracy. In addition, zircon is susceptible to problems of inheritance or the presence of xenocrystic grains derived from older source rocks or inclusions.

Over the past decade, the introduction of advanced analytical techniques such as high-precision, high-sensitivity mass spectrometry, low contamination chemistry, and ion microbeam technology has lead to an upgrading of classical geochronological techniques and the introduction of some new methods.

Development of the Sm-Nd whole-rock isochron method has made it possible to obtain ages on mafic rocks, which are often not datable by Rb-Sr or U-Pb methods and which commonly form the base of greenstone sequences (Hamilton *et al.*, 1979). The rare earth elements are considerably less mobile than rubidium or strontium, making the method less susceptible to open system behavior.

The introduction of the Ar/Ar method combined with step heating has led to a considerable improvement in the precision and reliability of ages compared to the K-Ar method. Ages obtained by this method for Archean rocks, however, still tend to be younger than those measured by Sm-Nd and zircon methods (Lopez Martinez *et al.*, 1984; Morrison *et al.*, 1985). Nevertheless, the method may hold considerable promise for dating later disturbances, especially if applied to individual minerals whose argon retentivity is well understood (Hanes *et al.*, 1985).

Recently a high-sensitivity ion microprobe capable of resolving mass interferences has been developed by a group led by Compston at the Australian National University. This instrument is capable of producing U/Pb isotopic analyses on polished zircon grains from spots 30 microns in diameter (Compston et al., 1986). Although the ages are at best of moderate precision, it is possible to date different generations of zircon growth within a single grain. Continued applications will determine the reliability of this method, but it promises to be of great value, especially in complex metamorphic terranes.

The high-precision U-Pb method was developed largely though the work of Krogh (1982). Since the author regards this as the most important method for detailed, accurate work, it will be discussed in some detail. The method involves selection of uncracked, clear zircon grains and removal of the outer rim of the crystals by abrasion. Apparently, much of the lead loss in zircon is confined to alteration along cracks and high uranium zones on the rims of zircon crystals. Selection of near-perfect grains and abrasion in most cases results in a reduced uranium level, almost negligible common lead, and considerable reduction of lead loss. This makes it possible to routinely measure ages to a level of precision of one part per thousand.

The stringent demand for sample quality in high-precision work makes it essential to have the capability of analyzing very small fractions. Enhanced ion transmission due to the use of extended focusing mass spectrometry permits the routine analysis of zircon fractions on the order of ten micrograms. With ion counting techniques it is possible to analyze single grains with only slight loss of precision. Analytical blanks of less than ten picograms are also necessary and achievable.

One of the principal limitations of the zircon method is the restricted range of rock compositions that contain low uranium zircon, generally intermediate and felsic rocks characteristic of later greenstone magmatism. Zircon is rare in the lower, largely mafic sequences. Baddeleyite (ZrO_2) gives reliable ages but this mineral occurs only rarely in some mafic rocks.

The reliability of the method for dating Archean rocks has been demonstrated by extensive application throughout the Superior province (see abstract by Davis et al.). The short time spans for some Archean magmatic cycles revealed by part of this work make it essential to apply high-precision methods to resolve geologic events.

Another result has been to show that many Rb-Sr whole-rock age determinations, formerly thought to be reliable, are on the order of 100 Ma too young, even on low-grade rocks such as late tectonic plutons (Birk and McNutt, 1981). The whole-rock isochron method involves a number of assumptions, such as closed system behavior, the assumption that all samples are cogenetic, and that they all had the same initial isotopic ratio. Whereas lack of collinearity of the data points indicates a violation of these assumptions, collinearity supports but does not prove them. Disturbance of the Rb-Sr ages may be due to a violation of the first assumption because of late movement of alkali-bearing fluids through the crust. This would accord with the observations by Corfu (1986), which suggest that during the late Archean the crust may have grown by a process of underplating so that deeper crustal levels are younger. Fluids released by this process may have disturbed Rb-Sr systems at higher crustal levels. Even some of the assumptions for Sm-

Nd whole-rock dating have been called into question in several cases where ages inconsistent with U-Pb measurements have been found (Cattell et al., 1984). Therefore, efforts to define orogenies largely on the basis of Rb-Sr whole-rock ages (Stockwell, 1982) may result in "hydrothermal" events recording metamorphic activity at the base of the crust and only indirectly related to the igneous and deformational history of the rocks themselves.

From the point of view of tectonics, the dating of deformational events and the precise characterization of the ages and metamorphic history of crustal terranes are of the greatest importance (e.g., see abstract by Hanson et al.). Deformational events can be bracketed by precise ages of samples, which can be shown on the basis of field relations and metamorphic texture to be older or younger than deformation. Thus, regional deformation in the Superior province has been shown to be a late event, following volcanism and lasting no more than about 30 Ma.

An idea of the thermal history of an area can be gained by analyzing more easily disturbed minerals such as sphene, rutile, and monazite. The sphene found in greenstone-associated rocks tends to give ages close to zircon while sphene from metaplutonic areas tends to be younger than zircon by several tens of million years (Corfu et al., 1985; Davis and Edwards, 1985). This indicates a slower rate of cooling for the metaplutonic areas, suggesting perhaps a sustained heat source beneath them. Much more needs to be understood about the blocking temperatures and conditions for metamorphic growth of these accessory minerals before their full potential can be realized.

The general absence of inheritance in zircons from greenstone-associated rocks has already been noted (see abstract by Davis et al.). This suggests that older continental crust was not involved in formation of the more fractionated greenstone belt lithologies, which may therefore have evolved in an oceanic environment. In contrast to greenstone belts, some high-grade metaplutonic terranes in the Superior province have been found to contain rocks with zircon populations of mixed age and to show widespread evidence for old sialic crust of pre-volcanic age.

A number of different possibilities can be envisioned for the tectonic relationship between the metaplutonic and the greenstone terranes. If the greenstones were floored by an older sialic basement, then evidence for inheritance in the zircon populations should be present throughout the evolution of the belt. If greenstones originated as rifts within an older continental terrane, inheritance may be present only in the earliest rocks, such as the lower bimodal, tholeiitic sequences. A late collision between an island arc and a continent might be expected to produce inheritance in the youngest igneous rocks.

The absence of inheritance is weak evidence, however, because older xenocrysts may have been selectively avoided during sample selection and because xenocrystic zircons may have dissolved in the melt (Harrison and Watson, 1983). A recent study using the ion microprobe to date zircon xenocrysts in basalts is a case where inheritance has been found (Compston et al., 1986).

Another way to study crustal contamination is to obtain reliable Sr, Nd, or Hf ratios on selected mineral grains from well dated rocks. Examples are the work of Hart and Brooks

(1977) and Machado (1985) on clinopyroxene, and Smith *et al.* (see abstract in this volume) on zircon. Model initial ratios obtained from whole rocks may, in some cases, be of questionable value in view of the problems with whole-rock ages.

Single grain analysis of zircons in the metasedimentary belts can be used to constrain the provenance and time of deposition of the sediments (Scharer and Allegre, 1982; Gariepy *et al.*, 1984). Present evidence indicates that deposition of some of these sediments in the Superior province may predate development of the late greenstone belts and they may be part of an early cratonic succession (see abstract by Percival and Sullivan; Wood *et al.*, 1986).

Geochronological techniques were originally developed to a large extent by physicists and chemists, many of whom had a limited understanding of geological complexities. Many geologists, on the other hand, have accepted ages without much consideration for their meaning, reliability, or even the quoted errors. It is important, therefore, to encourage an interdisciplinary approach to the problem of age dating. Geochronologists should have at least enough geological experience to follow proper sampling procedures while field geologists should be aware of the limitations inherent in any method. A "black box" or mechanized approach to the operation of geochronology labs may lead to the production of large amounts of dubious data, and a general cynicism in the geological community.

In conclusion, we have entered a new generation of geochronological and iosotpic research characterized by high-precision, high-sensitivity measurements on selected mineral phases, and a shared expertise between geologist and geochronologist. This will lead to the production of much stimulating data and is certain to modify our views of tectonic processes during the Archean.

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A Structural Viewpoint

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This workshop has been both significant and timely. The data presented has strongly underscored concepts and interpretations about the structure of greenstone belts that evolved in southern Africa in the period 1963 to 1974, and that involved the documentation of thrusting and fold nappes in the Archean.

There is a world-wide picture emerging of early recumbent isoclinal folds in many Archean greenstone belts, with associated imbrication and thrust faults, and involving some of the granitoids as well as the greenstones in places. At this meeting we have heard talk of cannibalistic nappes, stacking, sutures, and also klippe. Subsequent deformation, where present, is generally upright with a strong associated penetrative cleavage. The presented results of finite strain studies, both within and external to the greenstones, clearly indicate that such systematic studies provide interesting and elegant interpretations.

This meeting has clearly demonstrated that our resolution of the structure of greenstone belts has been improved because of the use of "high-resolution field geology": extensive, greenstone belt-scale, detailed mapping at 1:10,000 to 1:25,000 scale; Phanerozoic experience; quantitative finite strain mapping; careful use of structural facing—Shackleton's rule; more care with the use of field terms for lithologies; improved sedimentological and volcanological input; closely monitored geochronology and geochemistry; and recognition of tectonic contacts and tectonites.

The evidence for tectonic thickening and stratigraphic complication, despite the absence of fossils, is convincing. Fine tuning this, in order to unravel stratigraphic sequence, will be difficult and may be impossible in places. The use of detailed

sedimentological analysis and experimental igneous petrology could provide partial solutions to the fossil problem.

Our structural knowledge is essentially surficial, except in rare instances of exceptional vertical exposure, as in parts of Barberton. Many questions about the deeper structure of greenstone belts remain, with some geophysical studies providing interesting local models. The application of detailed shallow seismic reflection studies, with appropriate laboratory calibration of lithologies, may help detail some of this structure by defining the position and attitude of definable greenstone belt-scale contacts and discontinuities.

Further refinement of our knowledge of the structure of greenstone belts, and of their relationships to the granitoids, requires an integrated approach, in particular with much neglected metamorphic studies of textures and physical conditions; mineralization and alteration studies regarding the source, composition, and history of hydrothermal fluids; as well as a greater input of detailed sedimentology and precision geochronology.

Structural studies notable for their absence from the meeting were:

kinematic studies, especially of slides and faults;
incremental strain (path) studies;
speedometer (strain-rate) studies; and
the use of balanced cross-sections and palinspastic reconstructions.

The conveners have told me that I have a viewpoint to express. It is this: The evidence for the geological processes operative in the formation of greenstone belts is essentially structural. They are belts (as opposed to basins, rifts, bananas, platforms, or pumpkins) by virtue of their tectonic history. They owe their formation to deformation and deformation processes. If we are going to understand their origins, we must first understand their geometry and their kinematic evolution. Until we do that, we are simply pushing a viewpoint. Pushing a viewpoint is dangerous—it's like pushing a hearse; it invariably leads to dead ends. It is the interesting and stimulating, even controversial, interpretation that is more likely to lead to a reasonable answer. This does not mean having to be outrageous.

I remind you of what Tyndale-Biscoe (1949, p. 48) said of the Early Archean succession at Selukwe (Zimbabwe) when his mapping and observations of both sedimentary younging structures and conglomerate compositions (sedimentology) indicated an inverted and tectonic succession: "It is for this reason, mainly that the 'nappe' structure is invoked to explain the situation."

Subsequent more extensive studies in the same area led Stowe (1968) to write:

"The Allochthonous Nappe"

"The inverted Seluke Schist Belt slices, together with the schist and gneiss wedges in the Southern Gneissic Complex, constitute the lower imbricated limb of a large recumbent nappe fold. This rests on and was thrust over a stable gneissic basement represented by the Eastern and Western Gneissic Complexes. The contact between this basement and the schist cover is a wide shear zone, except in the east where the basement was reactivated and granite intruded the schist."

How could he be talking about the Archean? Surely this man was at the wrong conference! A member of the audience is quoted (Antrobus, 1968) as saying: "These new ideas of thrusting and nappes are startling."

Another, perhaps with acute hyperbole, remarked: "There is very little evidence of nappes at Barberton."

I wonder if there were any speakers at the wrong conference this time?

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An Ore Deposits Viewpoint

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A significant proportion of the contributions to the workshop stressed the perceived similarities between lithological successions, nomenclature, structural styles, and tectonic evolution affecting Archean supracrustal rocks and their younger counterparts. Several broad generalizations can be made concerning the differences and similarities between Archean and Phanerozoic supracrustal assemblages that place constraints on the tectonic processes resulting in formation of greenstone belts.

Volcanism. Archean volcanism is predominantly bimodal in nature (basalt-rhyolite), subalkaline, and subaqueous. There is a marked paucity of extrusive rocks whose volcanological, petrographical, and geochemical characteristics resemble orogenic andesites, although the term "calc-alkaline" is used frequently. The presence of extrusive ultramagnesium flows (greater than 20 wt% MgO) is unique to the Archean, although some magnesium picrites from Disko and Gorgona have some similarities.

Sedimentation. Archean sedimentary rocks are dominated by deep water volcaniclastic detritus, conveniently grouped as turbidites. Pettijohn's observation of the apparent lack of mature orthoquartzites and sedimentary carbonate rocks is still valid. The presence of vast quantities of banded iron-rich sedimentary rock of Archean age still has to be explained.

Metamorphism. In recent years, it has become apparent that vast regions of Archean supracrustal rocks (e.g., the Abitibi Belt of Canada) exhibit mineral assemblages characteristic of subgreenschist facies (Barrovian) metamorphism. It is of interest to note that metamorphic mineral assemblages found adjacent to some auriferous deposits are not consistent with metamorphic rank established by mineral assemblages on a regional basis. For example, felsic intrusions of overall granodioritic composition in the Timmins-Porcupine gold camp contain mineral assemblages ranging from quartz-albite-

paragonite-talc-stilpnomelane-chlorite-hematite to assemblages with quartz-paragonite/muscovite-chlorite-calcite-magnetite-chloritoid, suggesting a progression of alteration by oxidizing sodic fluids through to reduced alkali-depleted fluids.

By way of an introduction to the following comments, it may be appropriate to note that the majority of published and on-going studies of Archean geology have been carried out in areas spatially associated with ore deposits or in regions of anomalous metallic mineralization due to exposure, access, and the availability of a data base. This observation may have some important consequences regarding the applicability of regional syntheses. An ore deposit is the culmination of a fortunate sequence of natural processes that have concentrated a suite of elements by factors ranging from approximately 500× (Cu-Zn) to 5000× (Au) their average crustal abundances. The requisite fluid budgets and hydrothermal regimes responsible for these concentrations will have profound geochemical affects over areas several orders of magnitude greater than those of the deposits themselves. An ore deposit, by its nature, is genetically associated with one or several anomalies or complexities: stratigraphic, structural, mineralogical, geochem-ical, or geophysical—emphasis is placed on the term "anomaly," since geological monotony is anathema to mineral exploration. The point is that caution must be exercised in extrapolating observations made in "anomalous" areas to a synthesis involving much larger scales.

Little mention was made concerning fluid budgets and the scale of hydrothermal alteration in the Archean. As a first approximation, the masses of water contained in the Earth's three major reservoirs (the oceans, on and in the crust, and in the mantle) are approximately equal. Transference of water between these reservoirs over time may have some profound geochemical and isotopic consequences. Furthermore, the scale of hydrothermal regimes associated with igneous activity may not be appreciated, since the convective cooling of subaqeous igneous rocks may involve fluid volumes equal to magma volumes within several kilometers of the surface.

The study of Archean ore deposits has shown that rudimentary stratigraphy of greenstone belts can be recon-structed using the ore deposits. Greenstone belts containing volcanogenic massive sulphide deposits and stratabound gold deposits in talc-carbonate units are dominated by ultramafic rocks usually overlain by thin felsic and thick sedimentary rock sequences. Greenstone belts containing volcanogenic massive sulphide deposits are dominated by thick bimodal volcanic assemblages flanked by thick sequences of volcaniclastic sedimentary rock. The association of ore to specific lithofacies of greenstone belts reflects the broad tectonic environment active during ore formation. Also, the occurrence of gold but no base metal massive sulphide ore at Barberton, South Africa vs. occurrence of base metal massive sulphide but no gold ore at Manitouwadge, Ontario reflects fundamentally different processes responsible for the formation of these two greenstone belts. Economic geologists have long recognized the variability between greenstone belts and it would serve a useful purpose to better define and classify these belts according to their dominant rock types, stratigraphic sequence, structural style, and contained ore deposits.

The most striking difference in metallogeny between the Archean and younger times is the astounding accumulation of gold at the Earth's surface between 3.0 and 2.6 Ga. Nickel sulphide deposits and Algoma-type iron formations are also more common to Archean rocks than their younger counter-parts. Conversely, there are few notable examples of porphyry Cu-Mo deposits of Archean age, nor of Sn-granites, redbed Cu, Hg, or Mississippi Valley-type Pb-Zn deposits. Also, large oxide-facies banded iron formations of the Superior-type do not occur until Proterozoic time. Perhaps the most striking similarity in ore-deposit geology throughout time is the presence of volcanogenic massive sulphide Cu-Au deposits, which are broadly similar in most aspects, whether of Archean age or presently forming on the ocean floor. These metallogenic changes through time must be adequately explained by the tectonic evolution of greenstone belts through to Phanerozoic plate tectonics.

A number of participants referred to the primacy of field geology and field mapping in studies of the Archean. Maps are the lifeblood of mineral exploration. A few broad generalizations are offered: In mineral exploration maps are prepared rapidly, in a pragmatic fashion and using remote sensing as a primary source of information (particularly in regions of northern Canada where outcrop may average five percent). It may be said that the primary concern of map generation is consistency above accuracy and the recognition of patterns and complexities. An experienced mapper can cover approximately 1 square kilometer per week at a scale of 1:5000. With revisions at regular intervals as a result of joint ventures, restaking, compilations, and new ideas, the data base on file with many exploration companies in Canada is not only invaluable, but is also accessible.

A valid point to be made concerns the application of remote sensing to the rapid, accurate generation of maps of all scales. The availability and reasonable expense of airborne high-resolution cesium-vapour magnetometers and gradiometers, INPUT and airborne electromagnetic systems for scales greater than, say 1:20,000 and the reasonable cost of ground geophysics (for scales less than 1:5000, for example) suggests that this should be considered as a valid academic pursuit. Several thousand square kilometres could be flown with high-resolution magnetometer/gradiometer and electromagnetic systems for the cost of a moderate piece of analytical equipment. For smaller scales, where the emphasis is on structural complexities, the cutting of grid lines followed by ground geophysics could be done for a cost less than $8000 (1985) per square kilometer (and much less if done "in-house" through an applied geophysics program in a university).

The nature of Archean ore deposits was not emphasized during the workshop; however, much understanding of greenstone belts has come from study of ore deposits and their host rocks and the continuing exploitation of these ores provides significant impetus to further study and understand greenstone belts. Ore deposits in Phanerozoic and younger terrane exhibit typifying metallogenic signatures dependent on the geologic province in which they occur and the tectonic process that was causative to their formation. Similarly, Archean ore deposits provide a powerful tool to help unravel Archean tectonic evolution.

A Geophysical Viewpoint

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Little emphasis was placed at the workshop on the geophysical aspects of the study of Archean greenstone belts. This may reflect the common use of geophysics as an exploration tool but less common use as a research tool.

The focus of geophysical investigations has been to examine geometries and contact relationships and to look at the broader and deeper aspects of structure and interterrane relationships. Geophysical interpretations, primarily gravity models, consistently indicate that greenstones are restricted to the uppermost 10 km or so of crust. Gravity models suggest that granitic elements are similarly restricted. Seismic evidence demonstrates that steeply-dipping structure, which is characteristic of the belts at the surface, is not present in the underlying crust, which appears to have a simple layered structure. Geophysical evidence indicates that the boundaries between greenstone-granite and adjacent metasedimentary terranes are marked by large-scale crustal discontinuities. Within greenstone belts, measured stratigraphic thicknesses are often twice or more the vertical thickness determined from gravity modelling. This discrepancy may be explained if stratigraphy is repeated by thrust faulting and/or listric normal faulting. Where repetition is not a factor the gravity evidence points to the absence of the root zones of greenstone belts.

Geophysical methods are a complement to geological studies and are of particular use in areas of poor outcrop. As lithological units can have their own physical, electrical, and magnetic properties, it is possible to use geophysical methods also as a mapping tool. Their primary strength, however, lies in the information obtainable on the subsurface structures, particularly if geophysical models are constrained by integrating several methods.

Integrated geological and geophysical studies will undoubtedly provide considerable useful information on the nature of greenstone belts and help constrain models of origin and evolution. Given the increasingly available computing capacity to manipulate multiple data sets and sophisticated modelling algorithms, geophysics will play an important role in studies of greenstone belts and Archean geologic evolution in general.

A Thermal Viewpoint

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Two fundamental questions are raised when the thermal aspects of the tectonic evolution of greenstone belts are considered: (1) Was the average temperature of the Earth hotter in the Archean when most greenstone belts were formed? (2) Given that the modern lithosphere is laterally thermally heterogeneous, how relevant is the average temperature of the Earth to the local conditions of greenstone belt formation and preservation?

Intuitive logic suggests that the Earth had a maximum average temperature shortly after formation due to energy released by the accretion process, core segregation, and possibly short-lived unstable isotopes. Estimates of the global heat budget generally agree that of the order of 20% of modern global heat flow is derived from secular cooling of the Earth, the remainder coming from the decay of unstable isotopes of uranium, thorium, and potassium. If these estimates are correct, then the Earth had a higher heat loss and a higher average temperature during the Archean. If all modern heat flow is derived from internal isotopic heat generation, however, even though the rate of this heat generation must have been greater in the past, it is not *required* that the average temperature of the Earth was hotter. Rapid convection in the early Earth driven by heat of formation could have lost heat at a rate greater than it was produced by slow isotopic heat generation, resulting in a "cool" Earth prior to the establishment of the modern thermal regime of the Earth dominated by internal heat generation. Lunar crustal evolution suggests two phases of crustal magmatism, the early generation of an anorthosite crust followed by the "Archean" basaltic magmatism of the maria, perhaps representing a second peak in lunar temperatures due to radiogenic heating. Although the two-phase thermal evolution of the Earth may be considered unlikely, it must be remembered that with the present constraints on the modern thermal budget of the Earth, it cannot be assumed a *priori* that the Earth has cooled steadily throughout its 4.6 Ga history, although this assumption may be a useful working hypothesis.

More directly relevant to the question of the tectonic evolution of greenstone belts is the question "Was the Archean upper mantle hotter than the modern mantle?" It is probably a reasonable assumption that the average temperature of the Earth was hotter in the Archean than today, and that the global heat loss was greater, but again this does not *require* a hotter upper mantle if convective heat transfer was very efficient in the upper mantle. As the effective viscosities of mantle rocks are strongly dependent upon temperature, however, it is unlikely that upper mantle temperatures were significantly cooler in the Archean than today or convection would have been less efficient. By analogy with the Moon, up to about 100 Ma prior to the preservation of the oldest terrestrial rocks (3.8 Ga), Earth was subjected to a late global phase of impact tectonics (late heavy bombardment). If this impact activity had a significant heating effect on the outermost shell of the Earth, the time gap between the last phase of major impact activity on Earth and the preservation of the oldest rocks represents the time required to lose the impact heat from the Earth down to a depth of the order of 150 km. This scenario suggests that the pre-Archean upper mantle may have been rapidly cooling from an impact-heated phase.

These problems of the thermal evolution of the early Earth and implications for upper mantle temperatures were not discussed in detail at the workshop, but are relevant to early crustal genesis and greenstone belt evolution.

If it is assumed that global heat loss and upper mantle temperatures in the Archean were higher than the modern values, the question of lateral thermal heterogeneity must still be addressed with respect to the tectonic evolution of greenstone belts. Several mechanisms have been suggested for the mechanism of additional heat loss in the Archean such as faster/more sea-floor spreading, greater conductive heat loss,

and more intraplate (hot spot) heat loss. It appears that finding mechanisms for the additional heat loss is not a problem, but finding constraints for these mechanisms in the scant and often highly metamorphosed Archean rock record is difficult. Of fundamental importance to this problem is the question "How do Archean rock assemblages (including greenstone belts) differ from modern assemblages?" If there are no *significant* differences, then it is a reasonable assumption that the Archean tectonic processes differed only in rate from modern tectonic processes. If there are *significant* differences, however, the modification of tectonic processes by a hotter thermal regime must be considered (e.g., see abstract by Abbott and Hoffman).

Observational constraints on the Archean thermal regime are limited at present. Metamorphic gradients give valuable information about local thermal regimes during tectonic and magmatic activity, but must be treated with great caution in interpreting regional and upper mantle thermal conditions (e.g., see abstract by Morgan). The occurrence of komatiites in greenstone belts suggests high mantle temperatures for the generation of these magmas. However, discussion at the workshop indicates that more experimental studies are necessary to constrain the pressure and temperature conditions of the origin of these magmas before they can be used to tightly constrain Archean mantle temperatures. The question of lateral thermal heterogeneity must also be addressed in assessing the thermal implications of komatiitic magmatism—do the magmas indicate globally higher upper mantle temperatures or merely local hot spots in the mantle or vigorous eruption dynamics?

In contrast to the high mantle temperatures suggested by komatiites, diamonds of Archean age suggest that at least locally the lithosphere was cool and thick (similar to modern shield lithosphere). Our understanding of the thermal structure of modern shields (e.g., see abstract by Drury) does not preclude the local existence of thick lithosphere over a hot asthenosphere (e.g., see abstract by Morgan). Thick lithosphere may be preserved by balancing its basal temperature with the asthenosphere temperature by internal heat generation in a quasistable state: As the internal heat generation and/or asthenosphere temperature changes with time, minor changes in lithospheric thickness and/or metasomatic addition of more heat-producing isotopes may be required to maintain stability. Thus locally cold, thick Archean continental lithosphere may be compatible with hot mantle. However, if significant lateral variations in mantle temperature exist, cold, thick lithosphere may simply be developed over local cold spots (downwelling convection limbs) in the mantle with komatiites developed in the rising plumes. Furth ⌐ constraints on the longevity of thick lithosphere are required to resolve this problem.

Finally, in terms of global thermal conditions, great caution must be exercised in interpreting the remaining pieces of Archean crust as being fully representative of all Archean crust. Modern tectonic processes produce a great variety of sedimentary, igneous, and metamorphic terranes, and even today there is strong circumstantial evidence that some terranes are repeatedly reactivated while others are relatively stable. By analogy, not all Archean terranes are likely to have been uniformly preserved; in fact, the lack of high heat generation (high U, Th, K) terranes preserved in the Archean record suggests that these terranes may have been selectively reworked. A search must be made in younger terranes for reworked Archean terranes to attempt to piece together the full extent and diversity of Archean greenstone belts. Only when reasonable confidence is gained that the sample under study is not the result of selective preservation can the full implications of the thermal aspects of the tectonic evolution of greenstone belts be realized in terms of global tectonics.

A Sedimentological Viewpoint

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The study of Archean sediments gives direct access towards an understanding of Archean surface environments, the nature of the Archean surface, with its volcanoes, mountains, and basins, the evolution of the continental masses, and the early history of life.

At the workshop there was strong controversy marking a conflict of sedimentological and structural interpretations; and there was dispute among sedimentologists in attempts to probe deeply into the nature of the Archean surface. This account is a short summary of the workshop and an impression of the discussions.

The Tectonic Setting of Greenstone Belts

Rifting models. One of the most important applications of sedimentology to Archean rocks is in the study of the tectonic environment in which greenstone belts originated. Every greenstone belt is different, but many (though not all) appear to have originated in some sort of rifting environment. Facies analysis is helpful in understanding what happened. Some years ago, most sedimentary successions in greenstone belts were thought of as having been deposited in "basins," a very general term. More recently rifting models after McKenzie (1978) have been applied by McKenzie et al. (1980), Bickle and Eriksson (1982), and others to greenstone belts (Fig. 2). The model predicts specific sequences of sedimentary facies associations and volcanism during the extension of continental crust, and in many greenstone belts the "stratigraphic" succession appears to fit. Listric normal faulting is important in rifts, and has been suggested by some Archean geologists (including the present author) as one possible way out of the embarrassment caused by the exceedingly great apparent thicknesses of many Archean stratigraphic successions.

At the workshop, rifting models of various sorts were presented. Skulski et al., to cite one example, suggested that stretching by a factor of about 1.5:1 took place in La Grande River greenstone belt, Quebec, during a "classical" continental stretching event. Others saw rifting events in a variety of ways, and many authors discussed "pull-apart" basins or drew analogies with diverse modern settings. There was also considerable discussion of the contrast between "platform" and "rift" settings, after Groves and Batt (1984): It should be pointed out that "platforms" as used in this sense bear little resemblance to standard concepts of young sedimentary platforms. Most attendents at the workshop concurred that rifting might have been important in early stages of greenstone belt evolution but

SEA LEVEL

LITHOSPHERE

MAGMA
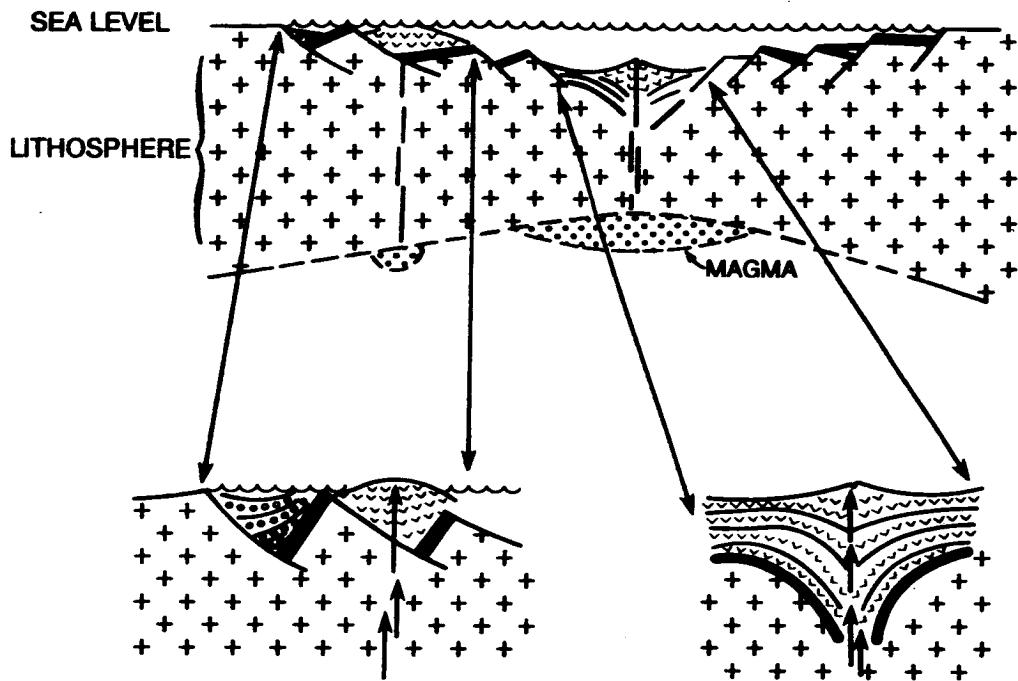
Fig. 2. Cartoon to show popular rifting models of the development of Archean greenstone belts. Top: cross-section of region of thinned crust; note listric faults. Solid black—sedimentary succession above unconformity; ooo—conglomerates, clastics; +++—lithosphere; vvv—volcanics; dots—magma. Bottom: details of two environments in top diagram. Left shows homoclinal succession unconformably on granite-gneiss crust; right shows synclinal succession dominated by lavas (for a rough modern analogue, see Mutter et al., 1984). Downwarping is produced by loading. Note that in both cases the first apparent folding is not produced by compression: In many greenstone belts the first deformation appears to have taken place prior to the imposition of a cleavage, which is often associated with a later compressional deformation. Later compression of the strata shown in the cartoon would generally tighten folds, increase dips to near-vertical and impose a cleavage (from Nisbet, 1986).

that evidence of rifting was usually destroyed in late events.

More generally, however, there remains the worry that the essential features of rifting models (such as the specific series of facies associations) have not been unequivocally established in any belt, nor have listric faults been proven. In many belts a very good case can be made to support a rifting model, though proof remains elusive. Perhaps with detailed paleocurrent work and precise zircon dating, a rift model may yet be proven: In the interim I strongly suspect that many of the Archean sedimentary/volcanic suites of greenstone belts, especially those that are homoclinal, formed in rift settings.

Ophiolites? Where rifting goes to infinity, new oceanic crust is formed. Four possible Archean ophiolites were described at the meeting. Here I mention sedimentological evidence. Strong controversy arose at the workshop about the Onverwacht "succession" of the Barberton Mountain Land. Does it represent something approximating oceanic crust, or is it a shallow-water succession of lavas and sediments? Among the most interesting of the discussions at the workshop was the debate between de Wit et al., who have used detailed mapping structural analysis to support the former notion, and Lowe and Byerly, who interpret sedimentological and stratigraphic data as suggesting the latter. De Wit et al. have built up an impressively argued case that the Onverwacht suite

includes a complex thrust belt, and that within this belt there exists a pseudostratigraphy comparable to that in young ophiolites; they have identified all the major components of an ophiolite complex. In contrast, Lowe and Byerly argue strongly that in the area they have studied (especially in the Fig Tree Group) the succession is thick and of dominantly shallow-water origin. In this argument, much depends on the interpretation of the depth of deposition of turbiditic sediments associated with the Barberton mafic-ultramafic rocks (Stanistreet et al., 1981; Lowe, 1982). The debate is important and constructive: with luck it will be resolved by more detailed fieldwork.

Unconformities. Ophiolites occur at or near the end of a rifting event; at the beginning there should be an unconformity. Archean unconformities are now known, and have been interpreted as demonstrating that certain (though not necessarily all) greenstone belts formed on continental crust. For instance, at the edge of the Belingwe belt, Zimbabwe, there is a very clear unconformity between 3.5-3.6 Ga tonalitic gneiss basement and 2.7 Ga greenstone (Bickle et al., 1975). However, there is no evidence to support the idea that this unconformity extends far beneath the belt. A second unconformity has also been inferred in this area, between 2.7 Ga greenstones and basement, but is not exposed. The Steep Rock Group, near

the edge of a late Archean greenstone belt at Atikokan in Northwest Ontario, comprises a suite of clastics, carbonates, and volcanic rocks laid down with clearly exposed unconformity on older basement (see abstract by Wilks and Nisbet).

However, all is not necessarily so simple. Hoffman discussed the significance of the Point Lake unconformity in the Slave province, Canada. He pointed out that the unconformity may not be a simple record of the initiation of subsidence in a basin or rift. Instead, it is possible that the conglomerate above the unconformity is the result of erosion during thrusting, as part of the sequence of events during the formation of a complex trench-arc system (see abstract by Lamb and Paris). In Hoffman's interpretation the unconformable relationship between volcanics and clastics represents a trench inner-slope setting, while subconformable relationships represent trench outer-slope settings. The greenstone belts of the Slave Province are thus seen as synformal remnants of a formerly continuous complex of tectonically accreted oceanic objects. Detailed sedimentological and structural fieldwork are likely to provide critical tests of this model.

Global Questions and Solutions Revealed Through Sediments

Continental growth. Underlying many of the workshop discussions was the more general question of how the continents grew, broke up, and aggregated again. There is clear evidence from several areas that very shallow-water sediments and evaporites are preserved among very old rocks [e.g., in the Pilbara (see abstract by Lowe and Byerly); in the Pongola belt (see abstract by Hunter *et al.*); and in Belingwe (Bickle *et al.*, 1975)] and that they apparently today overlie a normal thickness of continental crust. Since the thickness of continental crust depends on the depth of the oceans, this would suggest either that Archean oceans were—at least roughly—as deep as modern oceans or that substantial underplating of the continents has taken place in all areas where shallow water successions now overlie normal thickness of continental crust. On balance, there is an intuition (though not proof) that massive underplating of apparently stable cratons has not occurred: There is thus an opinion that the Archean oceans were deep. This idea is generally supported for a variety of reasons, not the least important being that it is difficult to see where the Earth's water would be stored if it was not at the surface.

The geochemistry of sediments has also played a role in the understanding of continental growth in the Archean: Whereas the mean age of the continents is between 2.0-2.5 Ga, it is still uncertain as to how much reprocessing and recycling of continental material has there been since the earliest recorded granitic rocks (ca. 3.8 Ga). Taylor and McLennan (1985) have used REE evidence from Archean sediments to suggest that a secular change took place around 2.7 Ga, and that in the early Archean continental volume (and hence area, if thickness has stayed moderately constant) was small. They have based their arguments on the REE geochemistry of fine-grained clastic sediments because these can be regarded as being representative of the bulk-composition of the upper crust. The Taylor/McLennan data consistently show negative anomalies in post-Archean samples, while Archean samples do not display Eu anomalies in their REE patterns. If the suite of sedimentary

rocks chosen is comprehensive, this would imply that sometime in the late Archean and early Proterozoic there was a major change in the nature of the continental crust, when a great volume of granitic material was freshly added to the continents from the mantle and then unroofed, to generate the Eu anomalies. Perhaps related to this postulated event is the often made observation that granites (sensu-stricto) are uncommon in the early Archean record: Most Archean granitoids are tonalitic and some field geologists are of the opinion, admittedly qualitative, that granites proper only become abundant in the Proterozoic and Phanerozoic.

At the workshop Naqvi (see abstract in this volume) presented very interesting results from the Chitradurga belt (2.6 Ga) in India. This belt contains a variety of greywackes and shales, most of which have been derived from surrounding tonalitic gneisses and also from K-rich granites. Rare earth element patterns in the greywackes generally show negative Eu anomalies. Perhaps the Chitradurga sediments represent a local and special case, but there is a strong suspicion that more general sampling of Archean shales will show widespread negative anomalies, and that there is not strong evidence in favor of a massive rapid change in the nature of the crust in the late Archean. Indeed, many sedimentologists hold the prejudice (admittedly unsubstantiated) that the abundance and distribution of the Archean cratons is such as would be expected, allowing for reworking, from a steady state model of continental volume (cf. Armstrong, 1981). Perhaps there was indeed a slow change in the composition of granitoid melts, as the average temperature of the mantle cooled, but more evidence is needed.

Archean life and the oxidation state of the early Earth. The geological record of life is mostly within the sediments of greenstone belts. There is excellent evidence for 2.7 Ga life in the stromatolites that are so abundant in the Late Archean (e.g., Belingwe, Fortescue, Steep Rock). In Barberton (de Wit *et al.*, 1982; Byerly *et al.*, 1986) and in the Pilbara (Buick *et al.*, 1981) there is now a small body of good evidence from stromatolites that shows that life existed by 3.6 Ga ago. Evidence from the analysis of RNA in modern bacteria and archaebacteria would suggest that both groups, including not only the ancestors of the cyanobacteria thought to have built stromatolites, but also the ancestors of archaebacteria now extant around mid-ocean ridge and terrestrial hydrothermal systems, are exceedingly old. Quite possibly the photosynthetic bacterial communities existing in shallow waters and now recorded in stromatolites represented only a part of the Archean living community: An equally important population may have flourished in and around Archean volcanoes, exploiting both ocean ridge and calc-alkaline heat sources. There may also have been an active population exploiting the photic zone in the oceans. Such a population is not recorded in the geological record, but it is reasonable to suppose that it existed, in much the same way that modern procaryote populations in lakes display an exquisitely stratified depth control. Living organisms breed explosively until all available resources are utilized, and it is thus reasonable to suppose that *if* life had managed to occupy shallow-water, shallow-levels in the oceans and also hydrothermal systems, then it would have done so rapidly in

every available place. The population of organisms may not have been diverse, but it would have been as dense as the sunlight, chemical supplies, and lack of predators allowed.

Finally, from the sedimentological record a reasonable, though not conclusive, case can now be made for the proposal that the atmosphere has been at least weakly oxidizing for most of the geological record. Sulphate evaporites in 3.6 Ga greenstones in the Pilbara and Barberton (Lowe and Byerly), and Nondweni (Hunter *et al.*, described at the workshop) lend some support to this notion.

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A Petrologic Viewpoint

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Instead of attempting to summarize what you all heard during the past two days, I will build on some of the topics and issues discussed to outline an approach to understanding the progressive development of continental crust in the Archean greenstone belts.

What interests petrologists is the relationship between the observed magmatic rocks, the inferred source rocks from which magmas are derived, and the processes inbetween. The possible processes are numerous, and they obscure the links between product and source. The processes are physical, and it is the physics of solid-melt-vapor systems that controls the chemistry of the magmas, and of the igneous rocks perserved in the geological record.

Using the Barberton Mountain Land as an example, the most prominent igneous rocks in early Archean terranes include (1) komatiites, (2) tholeiites, (3) tonalites and trondhjemites, grey gneisses, (4) potassic granites, pink gneisses, and (5) rare syenites. In some Archean terranes monzodiorites and syenites appear to correspond to the igneous stage represented generally by the grey tonalite and trondhjemite gneisses. Dacites and rhyolites may be abundant, presumably the surface expression of large magma chambers.

The possible source rocks are: (1) peridotite and eclogite of the mantle, (2) komatiites and tholeiites of the protocrust, possibly hydrated with formation of serpentine, talc, chlorite, epidote, and amphibole, (3) locally, garnet-amphibolite or amphibole-eclogite from oceanic crust thickened, foundered, or subducted, (4) tonalite gneiss of new continental crust, and (5) metamorphosed sediments buried by compressive tectonics, or subduction.

These observed magmatic products and inferred sources for the Archean are found in petrographic associations throughout geological history. The materials are the same, although additional materials become more important in post-Archean times. The one distinctive feature of the Archean is the existence of abundant komatiites, and this feature alone is sufficient to inform us that the asthenosphere was, at least locally, hotter by several hundred degrees than in later history. Two of the problems in komatiite petrogenesis are (1) attaining high enough temperatures, and (2) retaining the liquid at depth with its host peridotite for the relatively high percentage of melting required to generate the low-viscosity, high-MgO liquids. Because the thermal structure at depth is so fundamental for igneous processes, I maintain that the komatiites merit another round of detailed investigation, starting in the field, with petrography, mineralogy, and geochemistry; and with a major effort to strip off the effects of obvious and more subtle alteration, with experimental petrology of komatiites and peridotites to provide the calibration for the geophysical and thermal modelling.

More vigorous convection of the mantle in the Archean would probably be concentrated in plumes, and the existence of komatiites suggests that temperatures in these plumes may have been at least a couple of hundred degrees higher than in later regions of upwelling. The idea that pools of komatiite magma could be formed in the mantle in regions of local upwelling is very attractive, and its geophysical and petrological implications merit more attention. There is now persuasive, although not definitive, evidence that the density of komatiite liquids becomes higher than that of peridotite at a depth of 200–300 km. If so, komatiite magma formed at greater depths could not rise through this level and would in contrast have a tendency to sink. The conditions for intermittent release of komatiite from mantle magma chambers are speculative, but the tectonic conditions must involve tension. The reports of sheeted dikes in the Archean are probably more important as indicators of tension than as indicators of the possible presence of ophiolites that could be similar to modern oceanic crust.

MAGMAS \ SOURCES	PER.	ECLO.	Hb-GAB.	TONAL.	GRANITE
KOMATIITE	√	X	X —	X	X
BASALT	√	X	X	X	X
TONALITE	√ H_2O < 40 km	X	√ HIGH T	X	X
TRONDHJEMITE	X	X	√ H_2O √	X	X
GRANITE	X	X	X	√ H_2O √	
SYENITE	X	X	X	X	√ H_2O > 50 km

Fig. 3. Archean magmas: Possible and impossible sources. Based on major elements, mineralogy, and phase relations. Trace element and isotope geochemistry must also be considered.

If komatiites were derived from deep magma chambers, then the tholeiites can be interpreted as having been formed from a different source, lithospheric mantle at a shallower level, with heat provided by the deeper mantle plumes.

There is evidence for the storage of magmas of different composition in chambers at different depths. Calderas and associated volcanic activity confirm the presence of shallow silicic magma chambers. The widespread distribution of large anorthite crysts in many basaltic rocks is strong evidence for large, long-duration basaltic magma chambers at greater depths. In addition, large komatiite magma chambers may have existed in the upper mantle. There has even been discussion of a magma ocean, capped by lithosphere, but I prefer a picture with the chambers localized in regions of strong mantle upwelling.

Evidence from experimental petrology denies the prospect of deriving primary granitic magmas from normal mantle peridotite, and geochemical signatures leading to this conclusion must be satisfied by partial melting of young material derivative from mantle, such as basic rocks or greywackes. Tonalites and trondhjemites are derived not from the mantle, but from basic protocrust. For these magmas, we need additional experimental phase equilibrium data to define the ranges of pressure, temperature, and water content for their derivation. The structures of rocks in greenstone belts, leading to inferences about tectonic environment and process, need to be interpreted in terms of possible depths of formation and emplacement, for correlation with the experimental phase equilibrium data on the magmas. The coordination of these two approaches should lead to a clearer understanding of whether the granitoid magmas are formed as a result of crustal thickening, sinking of blocks of the crust, or an early version of subduction (presumably on smaller scales).

The general approach of using experimental petrology to unravel possible relationships between the observed magmatic rocks and the inferred source rocks is to follow the geochemists in "forward" and "inverse" approaches, and to use the phase diagrams to place major element constraints on the magic of minor element and isotope algebra. This approach neglects the very influential "processes" between source and near-surface products, but it provides a framework for starting to unravel the petrogenesis.

In the forward approach, the possible source rocks are subjected in the laboratory to variation in P, T, H_2O content and other variables, which provides specific information about the compositions of melts and coexisting minerals generated in the rocks under any conditions investigated. This sounds easy, but there are many experimental difficulties.

In the inverse approach, the near-liquidus phase relationships of an igneous product are determined through a range of P, T, and other variables such as H_2O content; the minerals on the liquidus of the particular composition must then correspond (in type and composition) to the residual minerals in a possible source rock at the specified conditions of pressure, temperature, water content, or other defined variables.

Much effort has been expended in these two approaches for peridotite-basalt, and less for peridotite-komatiite. Incomplete data are available for combinations of the series gabbro-tonalite-trondhjemite-granite-H_2O. On the basis of the available data, Fig. 3 is offered as a matrix of possible and impossible magmas from possible sources in the Archean. I

have assumed that in the Archean, deep subduction of cool oceanic lithosphere does not occur. I adopt the idea of a basic protocrust generated where tension permits uprise of magmas from mantle sources, followed by the formation of mini-continents, their migration and collision, with foundering of parts of the colliding continental nuclei, and local shallow subduction into an upper mantle hotter than it is today.

The sequence of igneous products in the Barberton Mountain Land, komatiites and basalt, tonalites and trondhjemites, granites, and finally syenites, also constitute possible magmatic sources. The sequential development of each magmatic product by partial fusion of the preceding igneous phase is consistent with major element phase relations. This interpretation appears to fit the physical conditions reasonably well, and appears to be reconcilable with much trace element and isotope geochemistry. Refinement of the structural geology and correlation with phase equilibrium results of experimental petrology might lead to a better definition of the extent of vertical movements in the Archean, and with the temperatures at various depths associated with these structural movements.

ABSTRACTS

HOT SPOT ABUNDANCE, RIDGE SUBDUCTION AND
THE EVOLUTION OF GREENSTONE BELTS

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A number of plate tectonic hypotheses have been proposed to explain the origin of Archaean and Phanerozoic greenstone/ophiolite terranes. In these models, ophiolites or greenstone belts represent the remnants of one or more of the following: island arcs (1,2), rifted continental margins (3), oceanic crustal sections (1,4), and hot spot volcanic products (1,3,5). If plate tectonics has been active since the creation of the earth, it is logical to suppose that the same types of tectonic processes which form present day ophiolites also formed Archaean greenstone belts. However, the relative importance of the various tectonic processes may well have been different.

The Archaean earth is postulated to have had greater internal heat production and consequently a younger maximum age of the oceanic lithosphere at subduction (6,7). One of the consequences of a greater proportion of subduction of young oceanic lithosphere in the Archaean is that ridge subduction would have been more common (7). The most common type of ridge subduction in the Archaean would have been that where oceanic lithosphere comprised both the overriding and subducting plate. The only present day example of this type of subduction is the subducting ridge in the Woodlark basin. This ridge crest has several geochemical anomalies: basalts with an island arc signature, and a dacite volcano on the ridge crest (8,9). The island arc component of the basalts has two proposed origins: contamination by an older subducting plate due to polarity reversal of the arc (9) and fluid contamination from the base of the subducting plate (10). Plate reorganization and ridge subduction are both postulated to have been more abundant in the Archaean (7). Regardless of the mechanism by which the arc-like component is generated, Archaean oceanic crust emplaced on land would have been much more likely to have an arc-like composition. Similarly, the dacite volcano observed on the Woodlark basin ridge crest could also have counterparts in Archaean greenstone belts.

Other aspects of the Woodlark basin subduction system may also have relevance for Archaean greenstone belts. The New Georgia island arc, which is being formed by subduction of the oceanic crust of the Woodlark basin (Figure 1), is composed of overlapping volcanoes, located 4-70 km above the Benioff zone (11,12). The New Georgia arc is quite different from a 'typical' Phanerozoic arc, e.g. the Marianas arc (Figure 2). In the Marianas, the volcanoes are spaced 50-100 km apart and sit 125-150 km above the Benioff zone (13,14). The island arc volcanics of the New Georgia arc also have some unusual characteristics. One island is a picritic volcano, thought to be the direct result of the ridge subduction process (8). If a higher percentage of Archaean island arcs were like the New Georgia islands, individual volcanoes would possess overlapping edifices and picritic volcanoes would occasionally occur. The overlapping volcanic edifices would increase the thickness of layer 2 (the pillow basalt layer) and would increase the probability of multiple phases of hydrothermal activity. Consequently, the relative abundance of Archaean ore deposits could be due to the greater incidence of New Georgia-like island arcs.

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Another probable consequence of greater internal heat production in the Archaean would have been a greater abundance of hot spot activity. For example, in the Phanerozoic, global ridge volume in the Cretaceous is thought to have increased and to have caused the Cretaceous sea-level high. This increase in the global sea floor creation rate may have coincided with an increase in hot spot activity (15). If increases in hot spot activity do coincide with increases in sea floor creation rate, hot spot activity must have been much more abundant in the Archaean. At present, 10% of all sea floor volcanism is estimated to result from hot spot activity (16). In the Archaean, it is likely that an even greater percentage of sea floor magmatism would have been hot spot generated.

Greater hot spot magmatism in the Archaean would have increased the incidence of bouyant subduction. Bouyant subduction can be a result of subduction of young oceanic crust or of older oceanic crust with a thickened crustal section (7). Much of the oceanic crust which subducts bouyantly has no volcanism or reduced volcanism. This reduction in volcanic activity as a result of bouyant subduction is most common if the overlying plate has a thickened crustal section. Consequently, an increase in hot spot activity in the Archaean could have decreased the percentage of subducting plates causing magmatic activity in the overriding plate, particularly when the overriding plate was relatively cold, thick continental lithosphere.

Areas of hot spot magmatism generally have a thickened pillow basalt section and a greater abundance of highly permeable rocks. These thickened pillowed sections can support more intense hydrothermal activity. Increased hydrothermal alteration at hot spots, particularly ridge-centered hot spots, could also have contributed to the relative abundance of Archaean massive sulfide deposits.

In conclusion, it is probable that many of the differences in preserved Archaean and Phanerozoic greenstone belt/ophiolite terranes can be explained as a result of a difference in the relative importance of different plate tectonic processes. This difference is a direct result of the increased internal heat production of the earth in the Archaean.

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HOT SPOTS AND RIDGE SUBDUCTION

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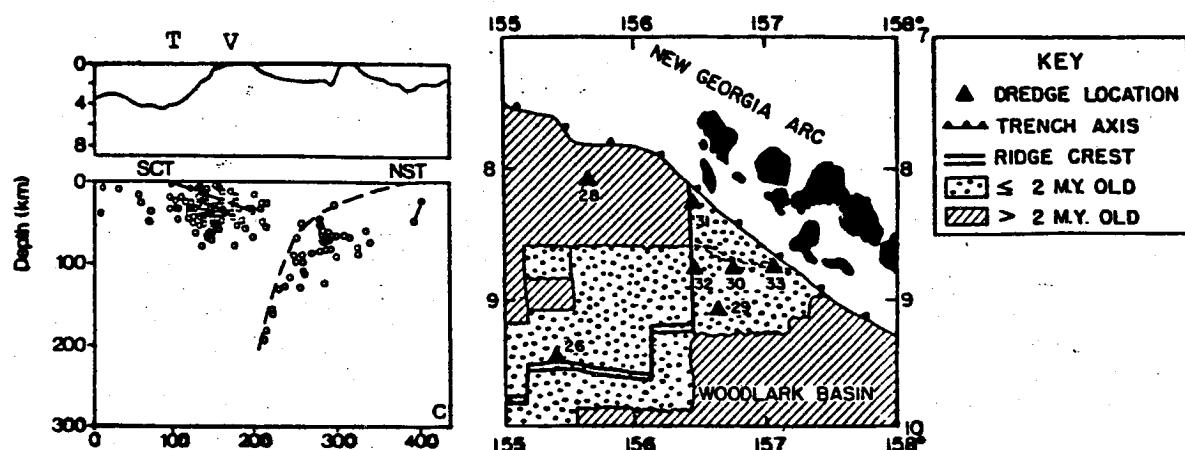
Figure 1. (left) Benioff zone of the New Georgia arc (SCT), after (12);
T = Trench, V = Volcanic Line.   (right) Volcanoes of the New Georgia arc,
after (10).
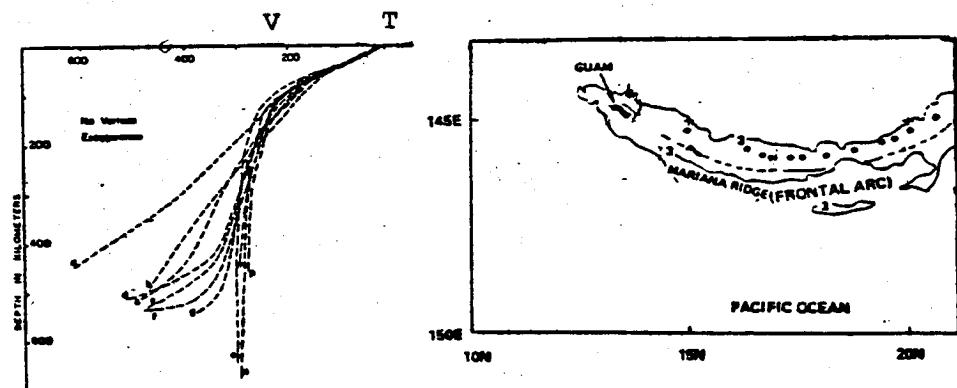
Figure 2. (left) Benioff zone of the Marianas arc, after (13); T and V as in
Figure 1. (right) Volcanoes of the Marianas arc are designated by dots,
after (14).

LITHOLOGY, AGE AND STRUCTURE OF EARLY PROTEROZOIC GREENSTONE BELTS, WEST AFRICAN SHIELD; Kodjo Attoh, Geology Department, Hope College, Holland, MI 49423

Distribution, Lithologic characteristics and Stratigraphic relations. Distribution of early Proterozoic volcanic rocks in the West African shield is shown in Fig. 1; an approximate boundary between Archean age terrane, to the west, and the Proterozoic terrane to the east, is partly marked by a major fault. Lithologic and chemical data have been compiled for belts (2-9) in the Proterozoic terrane from BRGM reports [1,2] . Available stratigraphic information from geologic maps of these areas indicate that a typical sequence is comprised of predominately mafic lava flows (basalt-andesite) at the base, which are overlain by felsic volcanic rocks including pyroclastic rocks and lavas. This succession, referred to as Lower Birimian, is overlain by Middle and Upper Birimian sedimentary rocks. Lithostratigraphic data from belt (1), located in northeastern Ghana [3] , indicate the volcanic succession is 6-8 km thick. The lowest unit in this succession is represented by 2 km of felsic pyroclastic rocks, flows and fine grained sediments. This is followed by 3-4 km of basaltic lava flows which are locally pillowed, the top of the unit is marked by a distinctive manganese formation (MF) consisting of Mn-Fe rich cherts up to 200 m thick. Dacitic lithic tuffs, welded tuffs and andesitic flows up to 2500 m thick overlie the mafic lava flow unit. The youngest volcanic unit consists of mafic tuffs and breccia with a distinctive fragmental texture. Preliminary data indicate that a similar succession occurs in belt (10). The internal plutonic rocks of belt (1) include: (a) hornblende-bearing granodiorite bodies considered to be subvolcanic plutons (σ-plutons); and (b) post-kinematic mica-bearing granitic plutons (pi-plutons). External plutonic rocks include tonalitic and granodioritic rocks which immediately flank the volcanic belt, and paragneisses which occur within the plutonic terrane.

Chemical charcteristics and Ages. Of about 100 chemical analyses reported for belts(2-9) calc-alkaline rocks consitute 55% and tholeiites 45%. Quartz-normative basalt constitutes 99% of the rock type in the tholeiitic suite. In the calc-alkaline suite, 9% of the analyses is basalt, 45% andesite and the rest is dacite and rhyodacite. The ratio of tholeiitic to calc-alkaline rocks based on the stratigraphic thicknesses and chemical analyses of samples from belt (1) is between 57% and 43%. Ultramafic volcanic rocks occur in belt (3), indicated from chemical data from belt (6) and (9) and constitute 1% of all samples analysed. Komatiites have not been reported from the West African Shield, thus the rocks analysed may be classified as high-Mg-basalts. The tholeiitic rocks from belt (1) are enriched in Ti, and depleted in Zr relative to modern ocean floor basalts [4] ,and are depleted in K, Rb, Sr and Ba relative to the calc-alkaline rocks. Within the calc-alkaline suite which include the subvolcanic plutons, the major and trace elements show continuous trends from calc-alkali basalts to rhyolites.

The hornblende-bearing plutons plot in granodiorite and diorite fields of Q-Kf-Pl diagram; whereas the rocks from the pi-plutons have normative and modal mineral compositions of granodiorite, quartz-monzonite and minor quartz synite and monzonite. All the plutonic rocks are strongly HREE depleted [6]. The -plutons (SiO2=56-66%) show the least depletion with $[La/Yb]_n$ = 13-43. The paragneisses of the external plutonic terrane (SiO2=70-71%) show the steepest REE pattern with $[La/Yb]_n$ = 33 - 66; while the post-kinematic plutonic rocks (SiO2=70-75), and La/Yb = 18 - 58, are somewhere in between. Relative to the subvolcanic plutons with (Th=1.9-5.7, and U=0.9 -1.9) the pi-plutons are enriched in Th and U (Th=7.7-10.9 and U=4.5-25ppm). Age of volcanism in the West African Shield is not known; however, K/Ar and Rb/Sr ages have been reported for the rocks which intrude the volcanic rocks and can be used to place minimum age limits. Rb/Sr analyses of mica pi-pluton samples from belts (2-9) yielded the following ages (my): 1870 ± 157 to 2004 ± 42 for whole rock; and 1940 ± 45 mineral (plagioclase) isochron [5]. K/Ar analyses of amphiboles from belt (1) gave the following ages: (i) an older age of 2223 ± 283 was obtained from hornblende in the youngest volcanic unit; and (ii) a younger age, 2087 ± 138 was obtained from zoned, titaniferous hornblende in a deformed diorite porphyry intruded into the lowest unit in the volcanic succession. The available data lead to the conclusion that the minimum age for the volcanic activity must be between 2200 and 2100 my. It is significant that Archean ages have not been reported from any of the volcanic belts (1-10).

Structure of an early Proterozoic Volcanic belt in northeastern Ghana. Cleavage in the volcanic belt strikes N40E and dips steeply to the NW and SE. Mesoscopic folds, with locally well developed axial surface cleavage parallel to this foliation, plunge steeply NW and SE. Because the orientation of fold axes and cleavage surfaces do not change with respect to the stratigraphic position, it is concluded that the whole volcanic succession was deformed during a pre-2000 my old orogenic event. Evidence for multiple deformation occurs in the form of NW plunging folds and the folded trace of the axial surface of the major folds. The strong NE-SW orientation of the major structures is such that one has to conclude that the second deformation was not as intense as the first. Foliation in the external plutonic terrane is subparallel to the foliation in adjacent volcanic rocks. Unequivocal evidence for pre-greenstone belt structure was not found in the external plutonic terrane; however, NS structures occur in the paragneisses, which are oblique to NNE-NE structures in the volcanic belt. Gravity anomalies associated with the greenstone belt and the internal plutons have been modelled taking the surrounding plutonic terrane as background. The model predicts that the depth to the bottom of the volcanic succession is 3-4 km. Fig 2 is a structural section of belt(1) based on gravity models especially with regard to allowable geometries of the rock units at depth. The overturned limb of the major anticlinal fold is consistent with available facing indicators.
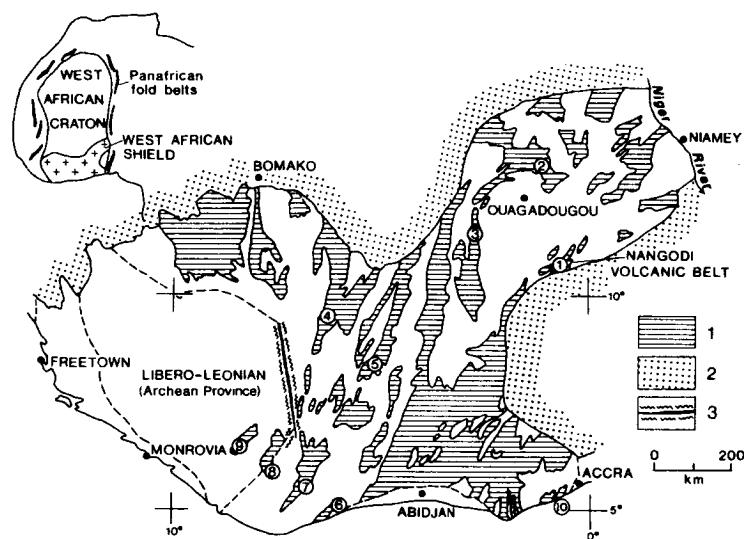
Fig 1. West African Shield showing the distribution of
Proterozoic volcanic-sedimentary belts: 1) early
Proterozoic volcanic belts, numerical labels referred to
in text; 2) late Proterozoic platform sediments; 3)
boundary between Archean and Proterozoic shields.
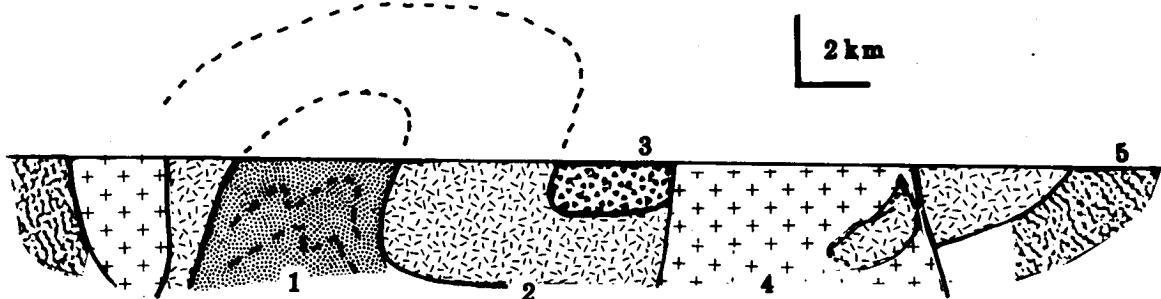
Fig 2. Geologic section across belt (1) in northeastern Ghana:
1) epiclastic sediments and tuffs; 2) mafic lavas
(tholeiitic basalts); 3) felsic tuffs and intermediate
lavas (calcalkaline); 4) postkinematic granites
(pi-granite); 5) granodiorites, tonalites and
paragneisses of external plutonic terrane.

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CORRELATIONS AND CONTRASTS IN STRUCTURAL HISTORY AND STYLE BETWEEN AN ARCHEAN GREENSTONE BELT AND ADJACENT GNEISS BELT, NE MINNESOTA;

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Introduction. Low-grade metagraywacke and greenstone of the Vermilion district and amphibolite facies schist and migmatites of the Vermilion Granitic complex (VGC) are separated by a series of east-trending dip-slip and strike-slip faults (Fig. 1)(1). Structural analysis in the boundary region between these two terranes indicates that they both sustained an early D_1 deformation which lead to recumbent folding. This was followed by a north-south transpression that resulted in the generation of upright F_2 folds and locally well-developed, dextral, D_2 shear zones (2). Despite these correlations, there are distinct differences in structural style and late-stage fold history between the two terranes that we attribute to: 1. differences in the crustal levels of the two terranes during deformation, and 2. effects of late-D_2 plutonism in the VGC.

D_2 deformation produced a series of upright F_2 folds with easterly striking axial planes that are the most prominent fold structures in both terranes. The largest fold of this series is a westerly plunging antiform that straddles the dip-slip fault boundary between the two terranes (Fig. 2). Large-scale parasitic folds on this structure are invariably of S symmetry in the southern VGC and occur on the northern limb of the antiform. D_2 dextral shear zones are well represented in the Vermilion district where they are generally parallel to the regional F_2 axial planes. Although distinct ductile shear zones are not observed in the VGC, evidence of a D_2 dextral shear component is locally indicated by asymmetrical pull-aparts and rotated vein segments in the migmatites.

F_1 recumbent folding is inferred from structural facing in the major F_2 antiform that crosses the boundary between the two terranes. Facing is downward on both limbs of the fold which is interpreted to be part of the lower, overturned limb of a large-scale F_1 recumbent fold (Fig. 2). A change to upward facing strata further south in the Vermilion district indicates a crossing onto the upper limb of this structure. Finite strain data, determined from clasts in sedimentary/volcaniclastic units in the Vermilion district, can be completely accounted for in terms of the deformation producing the F_2 folds (3). Locally intense F_1 folding in these rocks is therefore attributed to deformation in soft or very poorly lithified sediment. However, biotite schists making up part of the same structure in the VGC display a pronounced S_1 foliation that developed parallel to bedding during the early stages of metamorphism. We have suggested that metamorphic dehydration reactions occurring in the lower strata led to the development of high pore pressures in the upper portion of the sedimentary pile (4 and 5). The combination of high pore pressures and gravitational instability during the F_1 folding resulted in soft-sediment, coherent down-slope movement in the upper strata while the lower strata underwent strain and metamorphic recrystallization during F_1 folding. Soft-sediment F_1 folding in the Vermilion district could have led to a rather complex distribution of F_1 structures, because the more competent greenstones could not have been soft and therefore may have undergone a much different response to the F_1 folding.

F_3 folding has been observed only in the VGC to the north of the boundary zone with the Vermilion district, near the southwestern contact between the migmatites and the Lac La Croix Granite batholith. Along this

margin of the batholith, F_2 folds were reoriented during the emplacement of
the pluton and subsequently refolded by F_3 conical folds that formed during
the waning stages of the regional north-south transpression that generated the
F_2 folds (Fig. 3). Such F_3 folds are not observed along the southern margin
of the batholith where the F_2 folds are parallel to the batholith boundary and
therefore were not reoriented.

In summary, our analysis of the deformation along the boundary between
the Vermilion Granitic Complex and the Vermilion district indicates that the
two terranes have seen a similar deformation history since the earliest stages
of folding in the area. Despite this common history, variations in structural
style occur between the two terranes, such as the relative development of D_1
fabrics and D_2 shear zones, and these can be attributed to differences in the
crustal levels of the two terranes during the deformation. Similarly, the
local development of F_3 folds in the VGC, but not in the Vermilion district,
is interpreted to be a result of late-D_2 pluton emplacement which was not
significant at the level of exposure of the Vermilion district.

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EXPLANATION

Lac La Croix Granite
Wakemup Bay tonalite
Burntside tonalite
Amphibolite
Granite-rich migmatite
Schist-rich migmatite
Sedimentary & volcanic
rocks – Vermilion district

Vermilion Granitic Complex
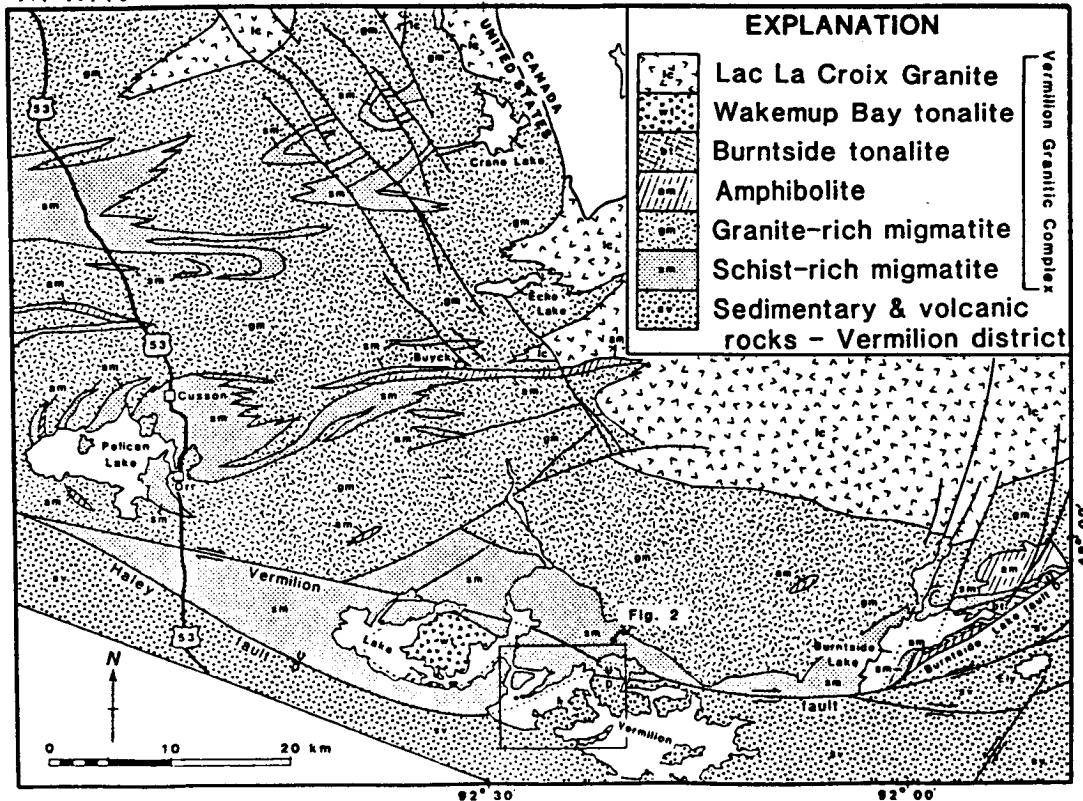
Figure 1. Geologic map of the southern Vermilion Granitic Complex and adjacent areas, after (6).

Fig. 3a. Major F_2 folds develop in both the western Vermilion district and the southern Vermilion Granitic Complex in response to a regional north-south compression. The Wakemup Bay pluton is emplaced during the later stages of F_2 folding and either emplacement of the pluton or emplacement of the Lac La Croix batholith reorients the adjacent F_2 antiform. At the same time, F_2 folds along the southwestern margin of the Lac La Croix Granite are reoriented by the rising batholith.
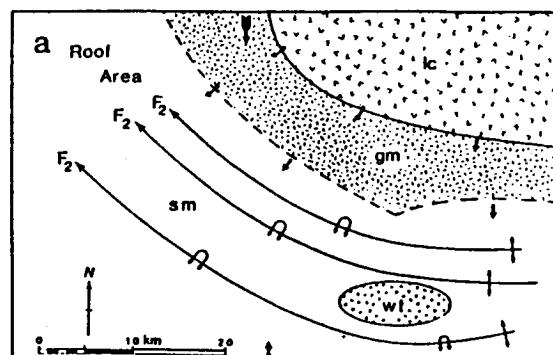
Fig. 3b. The reoriented F_2 folds adjacent to both the Wakemup Bay pluton and the Lac La Croix batholith undergo F_3 refolding as a result of continued north-south compression during the waning stages of M1 metamorphism. F_2 folds in the migmatites along the southern margin of the batholith undergo no significant reorientation during rise of the batholith, so that continued north-south compression results in continued flattening of the F_2 folds rather than refolding. The same continued flattening affects the F_2 folds in the western Vermilion district which is largely unaffected by local granitoid plutonism at the present level of exposure.
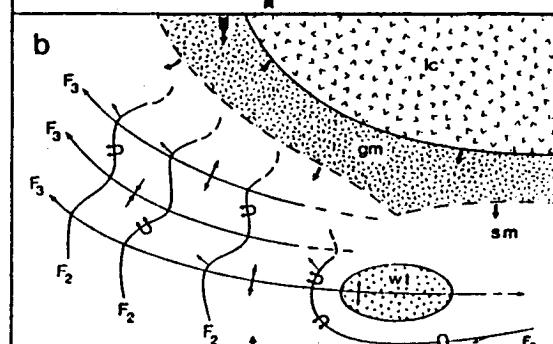
Fig. 3c. Following dip-slip displacement on the Haley fault, the F_3 fold cored by the Wakemup Bay pluton and the F_3 folds southwest of the Lac La Croix batholith are juxtaposed by right-lateral strike-slip displacement on the Vermilion fault.
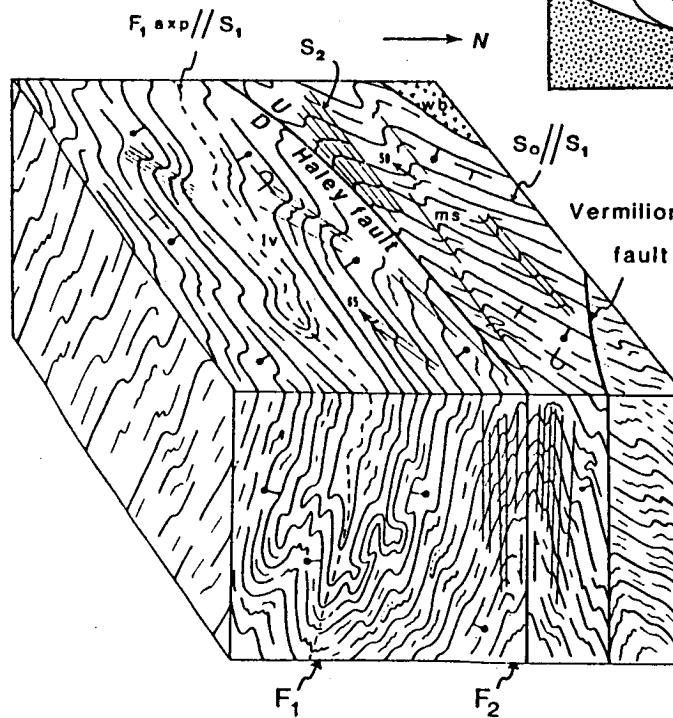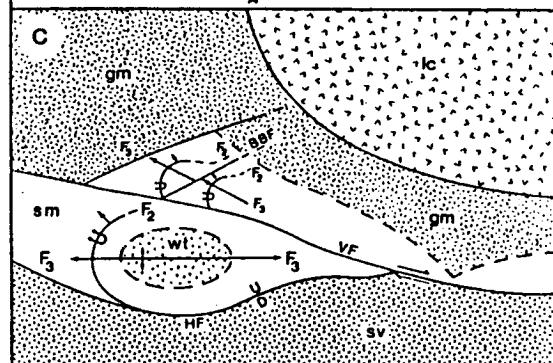
Figure 2. Schematic block diagram of area outlined in Figure 1 across Haley fault. Diagram illustrates variation in F_2 fold symmetry across Haley fault and relationship of F_2 to F_1 south of fault. Bar and ball symbols indicate local stratigraphic tops.

TECTONIC EVOLUTION OF GREENSTONE-GNEISS ASSOCIATION IN DHARWAR CRATON, SOUTH INDIA: PROBLEMS AND PERSPECTIVES FOR FUTURE RESEARCH

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The two fold stratigraphic subdivision of the Archean-Proterozoic greenstone-gneiss association of Dharwar craton into an older "Sargur group" (older than 2.9 Ga.) and a younger "Dharwar Supergroup" (1) serves as an apriori stratigraphic model. The concordannt greenstone (schist)-gneiss (Peninsular gneiss) relationships, ambiguities in stratigraphic correlations of the schist belts assigned to Sargur group and difficulties in deciphering the older gneiss units can be best appreciated if the Sargur group be regarded as a trimodal association of: (i) ultrabasic-mafic metavolcanics (including komatiites), (ii) clastic and nonclastic metasediments and paragneisses and (iii) mainly tonalite/trondhjemite gneisses and migmatites of diverse ages (2) which could be as old as c. 3.4 Ga. or even older. The extensive occurrence of this greenstone-gneiss complex is evident from recent mapping in many areas of central and southern Karnataka State.

The Dharwar Supergroup is deposited unconformably over an ensialic basement comprising the older trimodal association and is further divisible into a lower Bababudan and an upper Chitradurga groups. The volcanic and sedimentary rocks in the Dharwar schist belts display highly variable compositions, lithofacies and stratigraphic thicknesses. The available data is compatible with their deposition in a variably subsiding and progressively evolving basin(s) in an intracratonic or continental margin setting. The Bababudan group is dominated by sediments characteristic of the nearshore intratidal to shallow marine environments and subaerial toshallow marine volcanics (3, 4). The sediment thickness and way-up criteria are suggestive of progressive subsidence of the basin from south to north and concomitant accumulation of sediments derived from both intrabasinal and exterior sources which culminated in the deposition of thick (over 5 kms) sequence of polymict conglomerates and alluvial fan deposits in the rapidly subsiding Kaldurga basin (4). Subsequent sedimentation and volcanism proceeded in essentially deep marine environment as evident from rocks in the interiors of Shimoga and Chitradurga belts. The volcanic character evolved from predominantly tholeiitic (with minor komatiitic occurrences) in the lower units of Bababudan group to calc-alkalic affinities in the upper units of the Dharwar Supergroup. The overall major and trace element compositions of the Dharwar metavolcanics are comparable to Phanerozoic volcanics from continental margin or back-arc settings. While both light REE depleted and enriched types are noted often within the same volcanic formation, an important feature of the metavolcanics is their high Zr/Y character compared to most other Archean volcanic suites in the southern hemisphere suggesting possible trace element heterogeneities in the source regions of Dharwar volcanic rocks (5, 6).

The greenstone and gneiss formations throughout the craton show evidences of two or three phases of deformation with superposed folding resulting often in complex interference patterns. Both pre-Dharwar and Dharwar formations display broadly similar deformation styles and a remarkable parallelism in their tectonic fabrics differing in the intensities of deformation and grade of regional metamorphism (4, 7). The older sequences show superposition of tight upright or overturned isoclinal and/or recumbent folds of the first and second generations (F1 and F2) and a set of open folds (F3) and metamorphosed to amphibolite or granulite facies while the Dharwar rocks are generally in greenschist facies with large scale recumbent and tight isoclinal folds being uncommon (4). The structural history of the craton is complicated by repeated syn or late tectonic diapirism

and intense shearing, strike-oblique slip movements and thrusting particularly along several of the N-S trending regional shear systems (8).

Apart from the general problems concerning the conceptual approaches to early Archean tectonics and crustal evolution, the stages of the tectonic evolution in the Dharwar craton are poorly constrained by lack of information on many crucial aspects of the geology such as; chrono-stratigraphy of schist belts, timing of the major thermal and tectonic events, schist-gneiss relationships and their relative antiquities in the (older) trimodal association, the nature and evolution of the low grade-high grade transitions in the craton. Thus, while the evolution of the pre-Dharwar greenstone-gneiss association is largely enigmatic, the Dharwar Supergroup appears to be a consequence of wide-spread heating of the continental crust around c. 3.0 Ga., tectonic instability resulting in rifting probably along reactivated pre-existing lineaments, formation of broad basin(s), volcanism and sedimentation concomitant with variable rates of subsidence of the basin(s) in response to basement instability and differential upliftment of the surrounding basement highs (horsts?) across the boundary faults (4). The tectonic evolution of the pre-Dharwar crust and the relative importance of the "thick skin" vis-a-vis "thin skin" tectonics (4, 8) to the Archean/Proterozoic history of the Dharwar craton can be assessed only after more detailed structural data on a regional scale become available in conjunction with precise and reliable data on the primary and metamorphic ages of the schists and gneisses in the craton.

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GREENSTONE BELT TECTONICS - THERMAL CONSTRAINTS: M.J. Bickle (Dept
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Canada).

Archaean rocks provide a unique record of the early stages of evolution
of a planet. Their interpretation is frustrated by the probable unrepresenta-
tive nature of the preserved crust and by the well known ambiguities of tec-
tonic geological synthesis. Broad constraints can be placed on the tectonic
processes in the early earth from global scale modelling of thermal and chemi-
cal evolution of the earth and its hydrosphere and atmosphere. The Archaean
record is the main test of such models. It is the purpose of this contribu-
tion to outline what general model constraints are available on the global
tectonic setting within which Archaean crust evolved, and what direct evidence
the Archaean record provides on particularly the thermal state of the early
earth.
    The distinct tectonic style of Archaean granite-greenstone terrains un-
doubtedly reflects secular variation in the earth's tectonic processes as a
result of chemical and thermal evolution. Since tectonic processes are a
direct manifestation of heat loss processes in the earth, changes in the
earth's thermal state are likely to be primarily responsible for changes in
tectonic style. However, the geological record of tectonic processes is also
influenced by the state of chemical evolution of the solid earth and its
hydrosphere and atmosphere. As discussed below the basic volcanic dominated
nature of greenstone belts is probably as much a consequence of higher mantle
temperatures as any specific tectonic setting. Until proved otherwise we
must assume that 'greenstone belts' formed in as wide a range of tectonic
environments as modern sedimentary sequences. Care must be taken to distin-
guish features which are due to a specific tectonic environment from those
indicative of general tectonic processes in the Archaean earth.

Global Thermal Histories

    Calculations of global thermal evolution are based on derivations of
relationships between internal temperature and heat loss. Given such a rela-
tionship and the present temperature and radiogenic heat producing element
distribution within the earth it is possible to calculate temperature distri-
butions in the past with the assumption that the heat loss processes (convec-
tion) varied only in rate throughout earth history. Most current models are
formulated to satisfy the cosmochemical constraint that present day radiogenic
heat production produces about half of the total heat loss and that the earth
was hot soon after accretion [e.g. 1]. The main area of uncertainty intrinsic
in the modelling is the treatment of convection in a fluid of temperature
sensitive and non-Newtonian viscosity. One set of models, the 'parameterised'
convection calculations, derives a relationship between internal temperature
and heat loss by computing heat loss as a function of viscosity for a series
of models run with internally constant but differing viscosities and assuming
some form for the viscosity temperature dependence. Implicit in such model-
ling is the assumption that convection in a variable viscosity fluid can be
approximated by a constant viscosity appropriate to a characteristic tempera-
ture within the system. However, as first demonstrated by McKenzie and Weiss
[2] the assumptions of parametrical convection calculations are not approp-
riate to convection in variable viscosity fluids. Christensen [3] points out

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that it is the lower-temperature higher-viscosity upper boundary layer that
dominates convection rates and if heat loss should scale against any internal
temperature it will be a temperature within the upper boundary layer rather
less than the interior temperature (or correctly interior potential tempera-
ture which is mantle temperature normalised along an adiabatic gradient to
zero pressure). It is the interior temperature which is used for scaling by
the parameterised calculations. The difference may be illustrated by compari-
son of temperature - heat loss relationships.

The parameterised calculations lead to an expression for the relationship
between the Nusselt number Nu (the total heat flux to conducted heat flux) and
the Rayleigh number Ra of the form

$$Nu \, \alpha \, Ra^\beta$$

where β is between $\frac{1}{4}$ and 1/3. This relationship determines the temperature
sensitivity of the heat flux.

Christensen's calculations with variable-viscosity fluids suggest that
values of β around 0.05 are more appropriate over the limited range of the
experiments. The real uncertainties are rather greater than this given the
possibility of a layered mantle, two scales of convection in the upper mantle,
partition of heat loss between oceanic and continental regions and melting
with associated density changes with the upper thermal boundary layer. Sub-
stantial deviations in tectonic style from modern plate-tectonics could fur-
ther influence heat loss.

Two important conclusions may be deduced from the thermal modelling.
1. All the calculations indicate that the interior temperatures have not
changed by more than a few hundred degrees over most of Earth history, al-
though individual model predictions vary by a factor of two (Fig. 1).
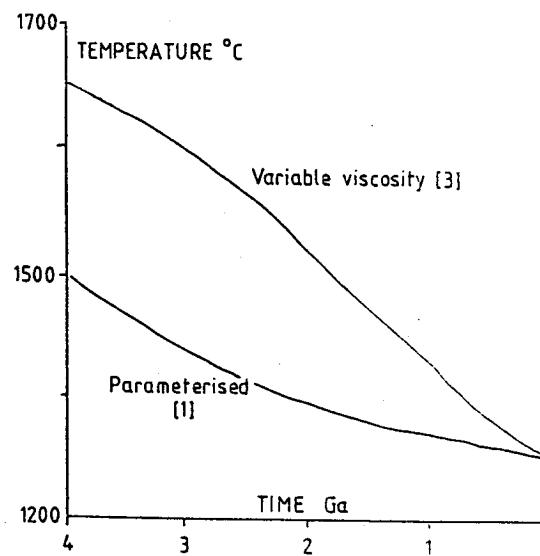
Figure 1. Comparison of variation of mantle temperature with time computed
for parameterised model [1] and variable viscosity model [3]. Models for
whole-mantle convection and approximately similar viscosity:temperature
functions.

2.   All the models predict that higher internal temperatures result in thinner, higher thermal gradient boundary layers (Plates)[1,3]. Further constraints must come from Archaean geology, which provides evidence on two critical parameters, upper mantle temperatures and continental lithospheric thermal gradients.

1.   Mantle Temperatures

The presence in Archaean greenstone belts of komatiitic lavas more magnesian than any younger lava is one of the few distinctive features of the Archaean and prime evidence that mantle temperatures were higher. To quantify the difference we need to know (1) the eruption temperature of komatiites and (2) the relationship between komatiite eruption temperatures and mantle temperatures. The first question has provoked surprisingly little discussion given its significance [e.g. 5,6]. Liquidus temperatures of komatiitic lavas are proportional to MgO content but this may be increased by olivine accumulation. Glassy, near phenocryst free lavas [7], and relict forsterite-rich olivine compositions have been taken to indicate liquids at least as magnesian as 27-30% MgO [5] although this is disputed [6]. Alternatively excess H_2O or alkalis have been suggested as fluxes lowering liquidus temperatures [e.g.8]. The latter is potentially testable through the temperature dependence of Ni olivine:liquid partition coefficients although such systematic tests have not been made. Even so eruption temperatures of ~1500°C (25% MgO) to ~1600°C (30% MgO) are 100-200°C hotter than any more recent lava.

The relationship between komatiite temperature and mantle temperature is more problematic. Adiabatically upwelling mantle cools along substantially higher thermal gradients (higher dT/dP) above the solidus as a result of the latent heat of melting (Fig. 2). If komatiites represent ~50% melts at high
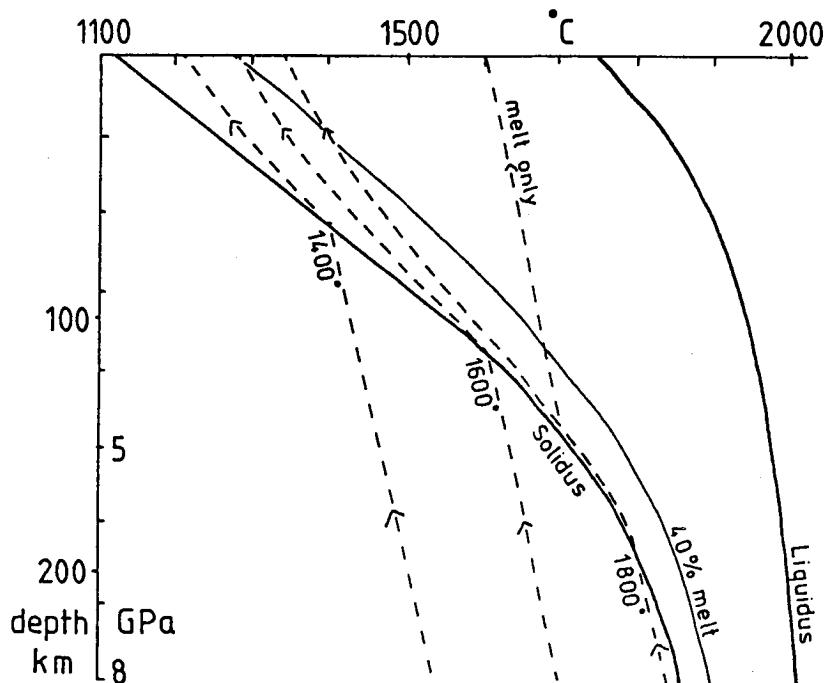
Figure 2.   Mantle liquidus and solidus and adiabatic ascent paths calculated with the assumption that melt and solid do not segregate on ascent, after McKenzie and Bickle [23].

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level with an olivine residue then a 1600°C komatiite must be derived from a
mantle in excess of 2000°C from depths >300km where we are essentially igno-
rant of mantle solidus-liquidus temepratures. Such temepratures substantially
exceed the upper bound of mantle temperatures derived from global thermal
modelling. Alternatively it has been suggested that eutectic melts at high
pressure shift to komatiite compositions [9]. Available phase equilibrium
data suggests this might be in the region 50-100 kbar [Fig.3]. If so koma-
tiites might be derived from mantle temperatures of 1800°C-1900°C, a potential
temperature of 1700-1800°C, and 400-500°C hotter than present day average
mantle. If komatiites are derived from anomalously hot upwelling convective
instabilities the potential temperatures of such regions are 200°C-300°C
hotter than mantle in present day thermal plumes.
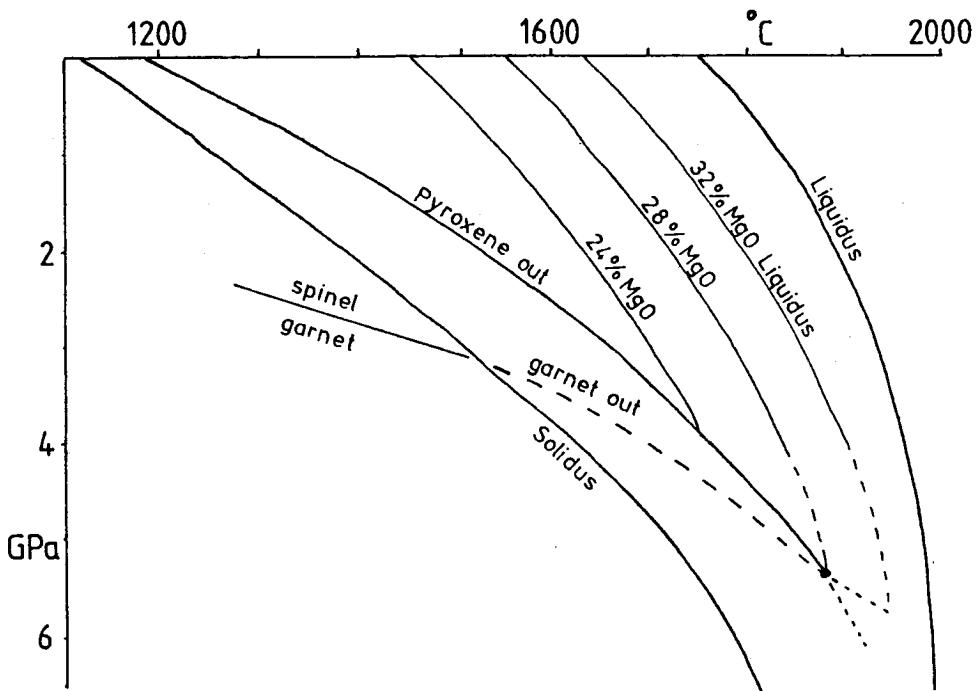
Figure 3. Phase relations for melting mantle-like compositions from
experiments on komatiites. Note intersection of garnet melting curve
with the pyroxene melting curve is hypothetical.

The chemistry of komatiites is not obviously reconcilable with their
being small degrees of eutectic melts. Incompatible element concentrations
are surprisingly uniform and are consistent with komatiites being ~50% melts
of plausible mantle materials [10,11]. Small degrees of melt would be
expected to be substantially enriched in incompatible elements although parti-
tion coefficients at the pressures of komatiite genesis are unknown and sub-
stantial modifications to komatiite chemistry by wall rock interaction might
be expected during their ascent [12].
Komatiite genesis is therefore problematic. However, even the most con-
servative estimates of komatiite eruption temperatures (a 25% MgO 1500°C lava)
implies mantle potential temperatures ~200°C hotter than at present and a 30%
MgO, 1600°C lava is inferred to imply mantle potential temperatures ~400°C
greater than today. One further complication is the possibility that at high
pressure the komatiite melt density exceeds that of solid mantle. If the in-
version in density is associated with a change in sign of the pressure

derivative of the potential temperature on the melting curve existence of a stable magma ocean at depth is probable [13]. The implications of such a magma ocean for global thermal and chemical evolution are profound.

2. Crustal Thermal Gradients

Metamorphic pressures and temperatures record anomalous thermal conditions in tectonically active crust. If sufficient is known about the tectonic setting of the metamorphism it is possible to invert the perturbed thermal conditions to infer steady state lithospheric thermal gradients [14]. Models for such inversion are mostly based on the thermal time constant over lithospheric thicknesses being rather greater than that of tectonic events (~<50 Ma). Given the possibility of magmatic or fluid heat transfer, such models tend to put upper bounds on lithospheric thermal gradients.

Archaean metamorphic conditions exhibit as wide a range of thermal gradients as modern orogenic provinces. High thermal gradients may at least locally be associated with magmatic advection of heat [15]. The lower thermal gradient, higher P/T metamorphism has attracted most interest as it places limits on the magnitude of lithospheric thermal gradients. The widespread 8-10 Kb, 700°C-900°C conditions recorded by gneiss terrains [16] imply background gradients little different from those in modern continental lithosphere. However, Morgan [17] suggests that these metamorphic conditions are buffered by crustal melting and heat flow in these regions is underestimated. Comparable high P/T metamorphism is known from upper-greenschist and amphibolite facies Archaean terrains [15,18-20] although it is less well documented. This is inconsistent with high heat flow through the underlying crust and not explicable as buffered by melting.

The inference from the metamorphic conditions of relatively low lithospheric thermal gradients has received substantial support from the observation of the formation and preservation of Archaean age diamonds [21]. These imply lithospheric thicknesses of ~150-200 km and mantle heat flux as low as 20 mWm^{-2}.

The observation that greenstone belts may have formed or been preserved in continental crust with relatively low thermal gradients has far-reaching implications for Archaean tectonics. Study of the metamorphism and its tectonic setting in greenstone belts would seem to be one rather neglected area of greenstone tectonics.

Implications on Global Thermal Evolution

The evidence for a significantly hotter mantle implied by komatiites is irreconcilable with the evidence for a thick cool continental lithosphere if the lithosphere behaved as its modern counterpart. There is good evidence from the depth-age relationships of oceanic lithosphere and sedimentary basin evolution that Phanerozoic oceanic and continental lithosphere behaves as a simple thermal boundary layer. To preserve a similar or greater thickness of Archaean lithosphere requires some additional process to stabilise the continental lithosphere. Morgan [17] suggests that increasing the concentration of radiogenic heat production might achieve this. It might but temperatures in the lower part of such lithosphere would be inconsistent with diamond stability at depths less than ~200 km.   An alternative mechanism is that the stabilisation results from density changes on melting [e.g. 22]. One consequence of a higher temperature mantle is that melting would start at much greater depths (Fig. 2)(~115 km for a 1600°C mantle versus

Bickle, M.J. and Nisbet, E.G.

~60 km for the present day ~1300°C mantle). The depleted zone is comparative-
ly less dense than unmelted mantle although whether the relatively small
changes are sufficient to stabilise the lithosphere against convective instab-
ilities is open to question. The mechanism of stabilisation of Archaean con-
tinental lithosphere and the formation and preservation of Archaean diamonds
is a key question. It has implications both for Archaean tectonic interpre-
tations as well as subsequent global evolution given the significance of the
continental lithosphere to continental tectonics.

There is one further significant tectonic implication of a hotter mantle.
The amount of melt produced by upwelling mantle is proportional to mantle
temperature [Fig. 4; 23]. With a 1600°C mantle any tectonic activity such as
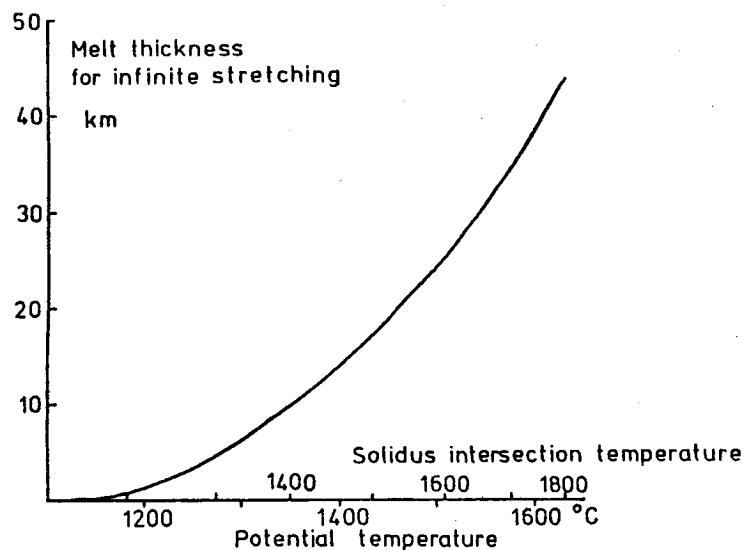
Figure 4. Melt thickness as a function of mantle temperature for infinite
           stretching (oceanic ridge case) after McKenzie & Bickle [23].

crustal extension which led to mantle upwelling would produce significant
magma. It seems probable that the basalt dominated nature of both Archaean
greenstone and late Archaean cratonic supracrustal sequences is a reflection
of mantle temperature and not necessarily of a special tectonic setting.
The extrusion of thick dense basaltic volcanics in supracrustal sequences
may be an important factor in the development of the characteristic tectonic
style of granite-greenstone terrains.

Archaean Tectonic Regimes

The prime assumption of all the global scale thermal models is that heat
loss processes changed only in rate. One hotly debated point is whether
plate tectonics or some alternative tectonic scheme operated during the
Archaean. For example, Richter [1] has suggested that once convecting mantle
penetrated the melt region below continental lithosphere the surface tectonic
regime would be dominated by vertical recycling rather than horizontal

motions. This scheme does not explain the preservation of the early Archaean crustal relics for which some special survival mechanism must be proposed. Perhaps the best evidence for major horizontal (plate) motions lies in the linear tectonic belts characteristic of the larger Archaean terrains (Superior Province, Yilgarn Block) and the evidence for large scale overthrust nappe tectonics in the high-grade gneiss belts. Other geological evidence is open to interpretation. For example, the significance of the calc-alkaline-like granite suites, possible analogies between some greenstone belt mafic sequences and ophiolites and the tectonic state of greenstone belts (allochthonous or authochthonous) are all disputed. One additional line of evidence does strongly suggest division of the Archaean earth into continental and oceanic-type regions. The heat loss through the Archaean continental regions inferred from metamorphic thermal gradients is too low by an order of magnitude to be representative of heat loss from the Archaean earth [24,25]. The extra heat is plausibly lost through oceanic-like regions as is the case today. This would involve substantial melting and recycling of volcanic crust.

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GREENSTONE BELTS ARE NOT INTRACONTINENTAL RIFTS. WHAT THEN ARE THEY?
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77058 and Geosciences, University of Houston, University Park. Celal Sengor
also Lunar and Planetary Institute, and Maden Fakultesi, ITU, Istanbul.

Hundreds of intracontinental rifts ("elongate depressions [within
continents] overlying places where the lithosphere has ruptured in extension"
ref. 1) with ages between 3.0 and 0 Ga have been recognized on earth (2,3,4).
Compressional features are either absent or insignificant in the vast majority
of these rifts. Prominent compressional features are reported from only a very
few rifts. (Notably: the Benue trough (5) the Dneipr-Donetz rift (Fig. 1) (6)
the Southern Oklahoma rift (7) and the rift occupying East Arm of Great Slave
Lake (8)).
Intense compression is the rule in greenstone belts and preservation of
regional extensional structures is rare. (Abstracts at this meeting). Whatever
greenstone belts are they do not satisfy the definition of intracontinental
rifts.
Wilson (9) showed that a common fate of intracontinental rifts is to
develop into oceans and that oceans are likely to close. Mountain belts mark
places where oceans have closed. In contrast to intracontinental rifts both
mountain belts and greenstone belts are dominated by compressional structures.
Pursuing Wilson's idea I therefore suggest that it might be useful for students
of greenstone belts to test the hypothesis that: "Greenstone belts are
mountain-belts marking places where OCEANS have closed". Ocean closing is a
complicated process (ref. 1) and some of the regional complexities that may be
recorded in greenstone belts are indicated in Fig. 2.
There is a possibility that students of greenstone belts are confusing each
other because some who describe greenstone belts as intracontinental rifts may
be consciously concentrating on an early episode in greenstone belt evolution
and recognize that the belts have a later oceanic and collisional history. I
suggest that this practise is confusing and is rather like describing Ronald
Reagan as a movie actor and ignoring more significant later episodes in his
career.

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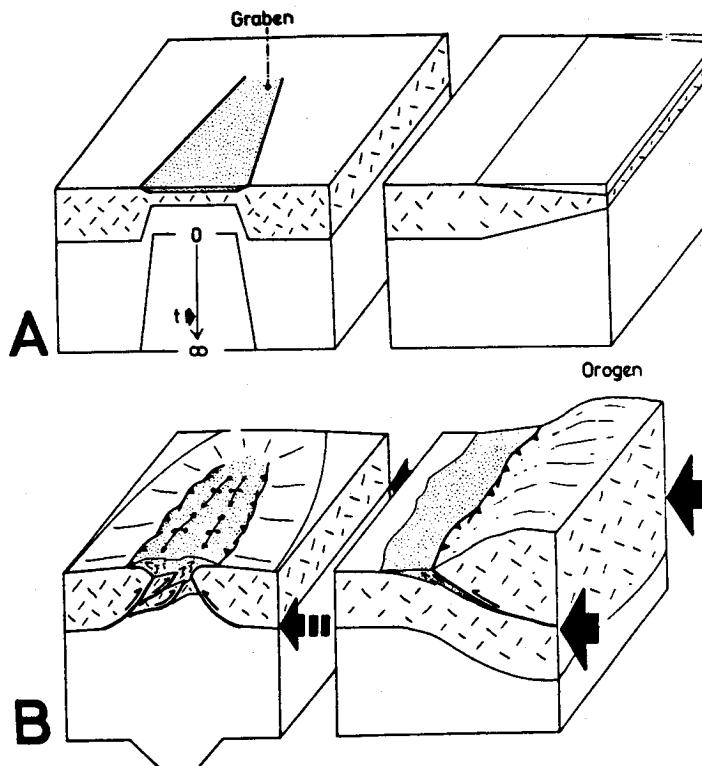
Figure 1. (from ref. 6) Illustration of how rifts within a continent
(such as the Dneipr-Donetz rift) have been affected by neighboring
continental collisions (as the Dneipr-Donetz structure responded to
collisions in North Dobrudja in the Early Jurassic). Observation
has shown that folding and thrusting developed in this environment is
much less intense than that with which we are becoming familiar in
greenstone belts.

GREENSTONE BELTS ARE <u>NOT</u> INTRACONTINENTAL RIFTS

Burke, K. and Sengor, C.
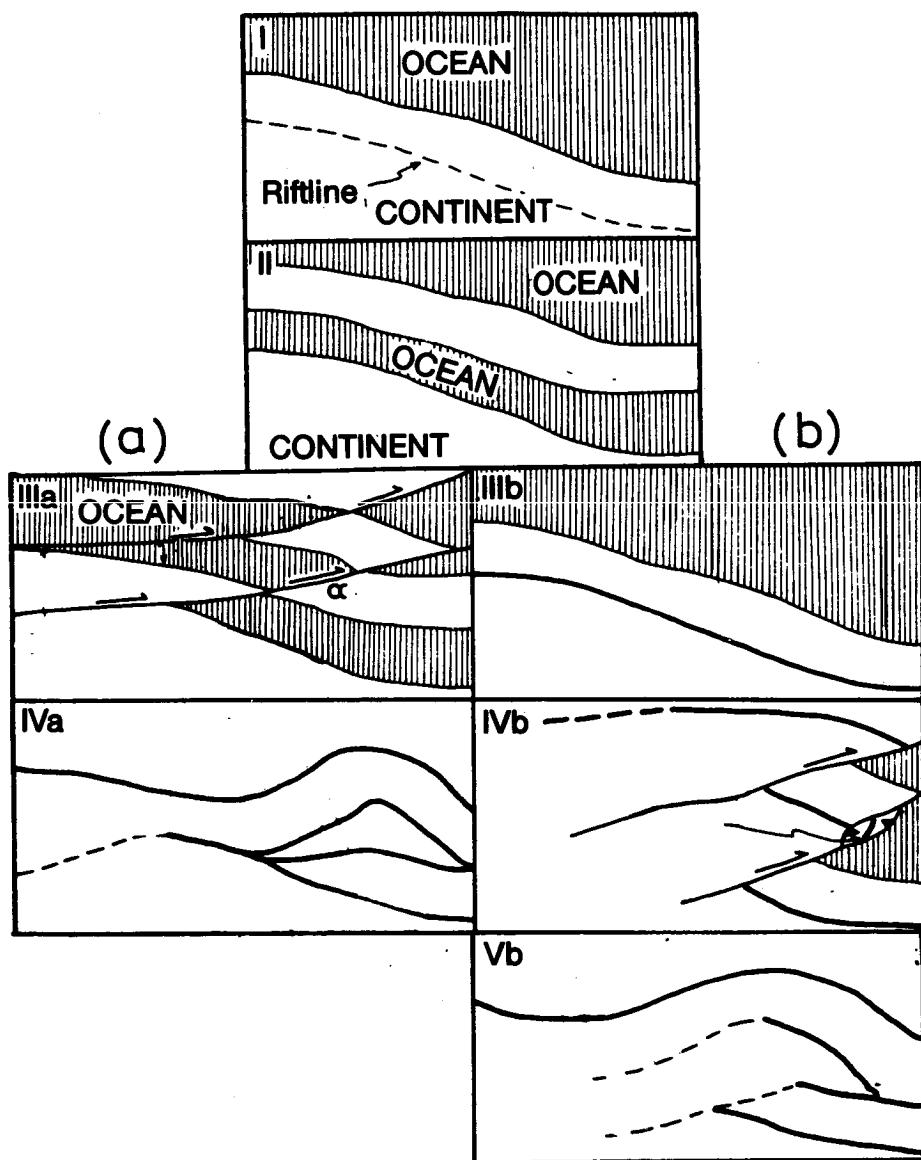
Figure 2. (from ref. 6) A possible origin for some greenstone belts. Rifting (I) takes a continental fragment out into an ocean (II). Major strike-slip motion (IIIa) is depicted as preceding collision between slivers of the continental fragment and the main continent (IVa). As an alternative suturing may take place (IIIb) before major strike-slip motion (IVb). In either case the preserved suture zones may end abruptly at strike-slip faults and late rotation may preserve puzzling polarities (Vb).

CAN TRACE ELEMENT DISTRIBUTIONS RECLAIM TECTONOMAGMATIC FACIES OF BASALTS IN GREENSTONE ASSEMBLAGES?

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During the past two decades many words have been written both for and against the hypothesis that the tectonic setting of a suite of igneous rocks is retained by the chemical variability within the suite. For example, Pearce and Cann (1) argued that diagrams can be constructed from modern/recent basalt subcompositions within the system Ti-Zr-Y-Nb-Sr such that tectonomagmatic settings can be reclaimed. If one accepts their general conclusion, it is tempting to inquire as to how far this hypothesis can be extended into other petrological realms. If chemical variations of metabasalts retain information relating to their genesis (tectonic setting), for example, this would be most helpful in reconstructing the history of basalts from greenstone belts.

Pearce and Cann (1) type diagrams are prepared by selecting a training set for which the tectonic settings of all of the analyses are known and obtaining a projection in which overlap of the fields of the known groups is minimized. IF, the training set is representative of a larger target population of interest, the projection may allow assignment of an "unknown" (an analysis not part of the training set) to one of the recognized groups. As the ratios of the variables are retained when percentages are formed, the search for such fields presumes that there are limits on the ratios of the three variables which identify a particular tectonomagmatic setting. The selection of three components and projection onto the plane of the ternary, however, does ignore potentially useful information and one could argue that a dimension-reducing procedure such as principal components analysis might lead to a more satisfactory and potentially useful display form.

However, a successful analysis of date with any multivariate procedure requires more than an understanding of the procedure itself. Additionally, the form of the data should be such that statistical procedures can be rationally interpreted. The subcomposition Ti-Zr-Y-Sr, for example, is part of a set of percentages and therefore subject to all of the concerns previously expressed by Chayes (2), Butler (3) and others concerning difficulties in interpreting both statistical measures of relationship (such as the correlation coefficient) and empathetic analysis of "patterns and trends" expressed in some compositional sub-space.

Simply stated, a set of composition percentages contains a mix of information from at least two sources:

    (1) physical/chemical relations among the variables
    (2) a change in the structure of the data as a result of
         a transformation such as percentage formation.

Statistical procedures typically allow one to recognize a behavior pattern that departs from a hypothesized expected behavior. The difficulty in interpreting percentages arises as a result of the mix of information noted above. For example, given

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a statistically significant correlation between Zr and Ti, can
one automatically assume that contribution from the mechanical
process of forming percentages is negligible? Is it possible, in
fact, that the mechanical process is the only one operative for a
given measured relationship? Can the investigator separate these
two effects in a given situation and assess their influence?

Until recently (Aitchison, 4,5) these questions received a
great deal of discussion and warning (Chayes, 2 and Butler, 3)
but no defined solution. Aitchison (4,5) presents a set of
procedures that ultimately are designed to allow an invstigator
to make use of information contained within a set of percentages
and these procedures are adequately described in the literature.
A training set of average Ti-Zr-Y-Sr analyses of 35 modern
basalts (including 24 from Pearce and Cann (1) and 11 drawn from
the current literature) with known tectonic settings was drawn
from the literature Space is insufficient to tabulate these raw
data and details will be published elsewhere; copies of the raw
data, however, are available from the author).

Aitchison's tests for basis independence (4) and complete
subcompositional independence (5) both reject their respective
null hypotheses (6). Thus, the investigator is assured that the
mechanical contribution is not dominant and that a physical-
chemical interpretation is warrented. Each analysis was
normalized to its geometric mean and eigenvalues and associated
eigenvectors extracted from the variance-covariance matrix of the
resulting log-row-centered data using principal components
analytical procedures. The first two eigenvalues account for
some 92% of the total variance and a plot of the first two
principal component scores is given in Figure 1. The boundaries
are empirical and constructed so as to isolate the known
tectonomagmatic groups. The distribution of scores successfully
delineates (1) the Within Plate Basalts, (2) the Ocean Floor
Basalts, and (3) the Arc Basalts. The principal component scores
are computed as follows:

Score 1 = -0.371*Ti-0.067*Zr-0.399*Y+0.836*Sr
Score 2 = -0.338*Ti-0.560*Zr+0.740*Y+0.158*Sr

where the individual variables are expressed in log-row-centered
form. In keeping with Pearce and Cann's suggestions (1), Ti is
defined as TiO_2 times 100 and Y is defined as 3Y. As one is
dealing with a logarithmic function, multiplication by a constant
changes the scale of the resulting projection but not the spatial
relationships. Ten sets of analyses from the literature were cast
into the space defined in Figure 1. In general, the
tectonomagmatic settings predicted from Figure 1 are in excellent
agreement with interpretations by the respective authors. Of
prime concern in this case, however, is the effect of
metamorphism on such subcompositions. Many authors (1) have
noted that Sr is easily mobilized during low to intermediate
grade metamorphism whereas Ti, Zr and Y remain relatively
constant. Three of the 35 analyses are plotted in Figure 2 with
additions and subtractions of 10% and 30% total Sr. These sets
of points define sets of straight lines which are subparallel to
the X-axis. Note that the "trend" of these lines is such that it

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may be possible to differentiate between Within Plate and Plate
Margin Basalts (Ocean Floor plus Arc) if the above model for Sr
mobilization holds.

  Perhaps a combination of detailed knowledge of the geology
of a particular greenstone assemblage plus judicious use of
diagrams analogous to Figure 1 will enable the investigator to
see through effects which heretofore may have masked
petrogenetically significant information.

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FIGURE 1. A plot of
the first two
principal components
for the training
set of 35 basalt
analyses. Numbers
refer to specific
analyses which are
available as a
separate from the
author. Boundary
curves are empirical
and drawn to
isolate the tectono-
magmatic facies.
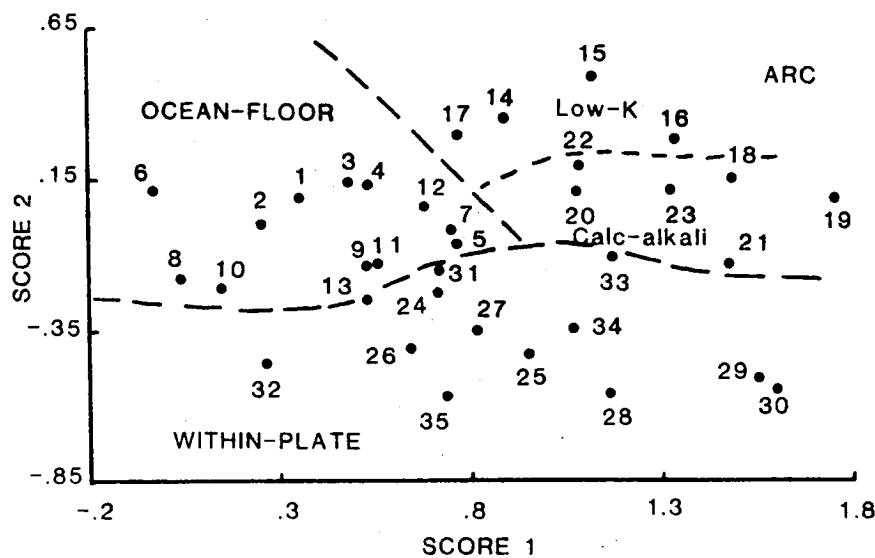
FIGURE 2. Additions
and subtractions of
10% and 30% Sr for
three of the basalt
analyses in the
space defined by the
first two principal
components. Note
that the trends are
parallel to each
other and sub-
parallel to the
boundary between the
Within Plate and
Plate Margin facies.
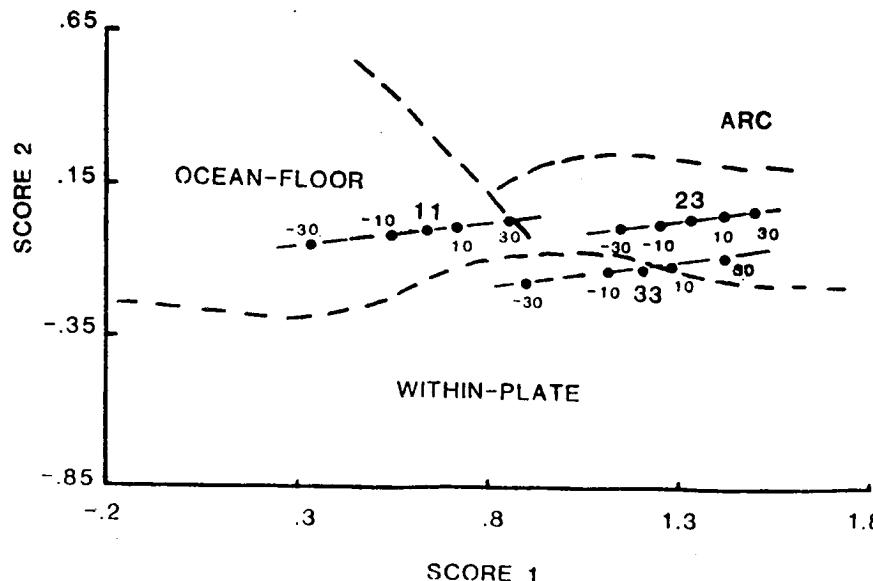
BARBERTON GREENSTONE BELT VOLCANISM: SUCCESSION, STYLE, AND PETROGENESIS; Gary R. Byerly and Donald R. Lowe, Department of Geology, Louisiana State University, Baton Rouge, Louisiana 70803 USA

The Barberton Mountain Land is a small but remarkably well-preserved and accessable early Archean greenstone belt along the eastern margin of the Kaapvaal Craton of southern Africa. Although there is some question about the role of structural repetition of various units, detailed mapping in the southern portion of the belt leads to the conclusion that a substantial thickness is due to original deposition of volcanics and sediments (1). In the area mapped, a minimum thickness of 12km of predominantly mafic and ultramafic volcanics comprise the Komati, Hooggenoeg, and Kromberg Formations of the Onverwacht Group, and at least one km of predominantly pyroclastic and epiclastic sediments derived from dacitic volcanics comprise the Fig Tree Group. Much greater apparent thicknesses of Fig Tree are due to numerous structural repetitions. The essentially non-volcanic Moodies Group lies conformably on top of the Fig Tree. The position or correlation for the Sandspruit and Theespruit Formations relative to the above units is not known. The Barberton greenstone belt formed primarily by ultramafic to mafic volcanism on a shallow marine platform which underwent little or no concurrent extension. Vents for this igneous activity were probably of the non-constructional fissure type. Dacitic volcanism occurred thoughout the sequence in minor amounts. Large, constructional vent complexes were formed, and explosive eruptions widely dispersed pyroclastic debris. Only in the final stages of evolution of the belt did significant thrust-faulting occur, generally after, though perhaps overlapping with, the final stage of dacitic igneous activity.

The volcanic succession in the Barberton greenstone belt is often used as a general model for greenstone belt stratigraphy (2). Previous workers have suggested that volcanism there was cyclic, ultramafic to mafic to felsic, on a scale that ranged from tens of meters to tens of kilometers in stratigraphic thickness, with small cycles superimposed on large cycles. In the grossest sense, the base of the sequence is predominantly komatiitic and the top dacitic, but beyond this the detail of volcanic succession is complex. Thin units of dacitic tuff are recognized within the Komati Formation and komatiitic lavas are interbedded with Fig Tree Group sediments. Simple, small-scale cycles are not present. Sequences previously interpreted as small-scale cycles are now known to represent thick, stratiform alteration zones of mafic and ultramafic lavas to silicic rocks with a remarkably calc-alkaline-like chemistry (3). Systematic increases in Si,K, and Rb accompany decreases in Fe,Mg, and Ni, while Al,Ti, and Zr remain constant from base to top in these cyclic units(4). Throughout these alteration zones the flows typically have mafic volcanic textures and structures, and are usually fine-grained and in places pillowed. Preserved spinels in silicified rocks initially crystallized in mafic or ultramafic lavas. After taking into account the nature of this common style of alteration it appears there are no obvious systematic trends in lava composition in the stratigraphic sequence. Notably, however, the two thick sequences of dacitic volcanics seem to represent prolonged volcanic episodes with no mafic or ultramafic activity.

Komatiitic or basaltic igneous activity seems also to occur with little or no other type of igneous activity in three or four thick sequences.

Styles of igneous activity vary primarily as a function of lava composition. Komatiites throughout the sequence occur as massive flows with typical spinifex textures or as thick flows that often display cumulus-layered bases or as pillowed flows and only rarely as hyaloclastite units. In most sections the flows are quite thin, typically 50cm to 5m, and only rarely up to 50m. They are rarely vesicular, suggesting deep water extrusion, but in several sections interbedded sediments are of shallow-water origin. We have observed no vent complexes for the komatiites, though they are relatively widespread units. The komatiitic unit within the Hooggenoeg can be traced for over 50 km of strike around the Onverwacht anticline. The komatiitic unit beneath the Msauli Chert crops out over a similar distance. Only in the uppermost komatiitic unit is there a local lateral facies; here the lavas interfinger with shallow marine sediments and were of more local extent, though again no vents are recognized. Basaltic igneous activity is characterized by thick to thin lava flows, in places pillowed or massive and only rarely by pillow breccias. Two separate basaltic sequences in the Hooggenoeg crop out for 50 km along the Onverwacht anticline. These lavas are non-scoriaceous, but commonly contain up to 5% vesicles primarily as radial vesicles about the margins of pillows. The lower basalts of the Kromberg occur as a thick sequence of lapilli tuffs, especially thick on the west limb of the Onverwacht anticline (5). These units are locally crosscut by irregular dikes and sills of basalt, and in places contain blocks and bombs of both juvenile and accidental lithologies. They appear to represent near-vent facies and were perhaps similar to modern basaltic cindercone fields. Some lithologies in this unit are moderately scoriaceous. Laterally, these units are represented by interbedded sediments and pillowed to massive lava flows. Dacitic igneous activity is represented on two different scales: by the relatively common tuffaceous units that occur throughout the section, and by very thick sequences of lavas, pyroclastics, and epiclastics at two locations in the sequence (1). Thin, typically a few tens of cm but rarely to a few tens of m, tuffaceous units occur throughout the sequence, and are usually completely altered to a micromosaic of quartz and sericite. Textures are remarkably well preserved, however, and indicate highly pumiceous particles often in the form of accretionary lapilli commonly in graded airfall beds. These units are regionally extensive, greater than 50 km strike length, but associated vent complexes are not found. The two major dacitic units, at the top of the Hooggenoeg and top of the Fig Tree, clearly represent vent complexes. They form complex associations of lava flows or domes, breccias, and tuffs hundreds of m thick. Along strike systematic changes in lithologies can be recognized where sedimentary rocks represent debris being shed off the constructional vent complex. These units do not appear to be laterally interbedded with more mafic lavas.

Petrogenesis of Barberton greenstone belt volcanics is not likely a single, one-stage process. Indeed, the succession of units and common isolation of one compositional group from the others may even

require a separate petrogenesis for komatiites, basalts, and dacites. Komatiites from the top and the base of the sequence are remarkably similar in composition (4,6,7). They are typical of komatiites worldwide except for very low Al/Ti, very high Ti/V, and other ratios that require a very depleted upper mantle source (6). Otherwise, most compositional variation within the komatiitic suite seems consistent with low-pressure fractionation of olivine, later joined by clinopyroxene in komatiitic basalts. Basalts of the Hooggenoeg and Kromberg Formations have typical tholeiitic compositions, including a pronounced iron-enrichment and lack of alumina-enrichment, that can be produced by low-pressure fractionation of plagioclase, clinopyroxene, and olivine. Immobile trace elements and their ratios, such as very low LREE/HREE, also require a depleted upper mantle source. Compositional data are not inconsistent with a single liquid line of descent of komatiites and basalts. While both komatiite and basalt sequences suggest substantial low-pressure fractionation there is not generally an adequate mass of layered intrusives to account for this fractionation in situ. The Barberton sequence contains less than 5% layered intrusives, yet basaltic komatiites and Fe-rich basalts each require 50% or more fractional removal of crystalline phases from their parental melts which must have taken place beneath the present level of the greenstone belt. Dacites are the only intermediate to silicic magmatic rocks found. They range from 60-70% SiO_2, 15-16% Al_2O_3, and have Na_2O/K_2O ratios of about 3 in the freshest samples and are thus trondjhemitic in character. They display extreme fractionation of LREE to Y, and have very high concentrations of highly incompatible elements such as U and Th. Plagioclase and hornblende are the major phenocryst phases in all dacites. Some also contain either quartz or biotite as phenocrysts. Their compositions suggest a source that was mafic in composition and a relatively small degree of partial melting of an assemblage dominated by amphibole. They are not related to associated basalts by any simple, one-stage magmatic process, though could be related to a second stage of igneous activity at the base of a thick, hydrated pile of mafic volcanics.

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TECTONIC SETTING AND EVOLUTION OF LATE ARCHEAN GREENSTONE BELTS OF SUPERIOR PROVINCE, CANADA: K.D. Card, Geological Survey of Canada

Late Archean (3.0-2.5 Ga) greenstone belts are a major component of the Superior Province of the Canadian Shield where alternating, metavolcanic - rich and metasedimentary - rich subprovinces form a prominent central striped region bordered in part by high-grade gneiss subprovinces, the Pikiwitonei and Minto in the north, and the Minnesota River Valley in the south. The high-grade gneiss subprovinces are characterized by granulite facies gneiss of plutonic and supracrustal origin, and by abundant plutonic rocks. Minnesota River Valley has rocks older than 3.5 Ga; absolute ages of Pikiwitonei and Minto rocks are unknown but Minto does have north-south structural trends distinctive from the dominant east-west structures of Superior Province.

LEGEND
⌐⌐ Proterozoic, Phanerozoic rocks
— Subprovince boundary
ARCHEAN SUBPROVINCE TYPE
Plutonic
Volcano-plutonic
Metasedimentary
High-grade gneiss
300 km

Volcano-plutonic subprovinces of Superior Province consist of generally narrow sinous greenstone belts bordered and intruded by voluminous plutonic rocks, including tonalitic gneiss, synvolcanic plutons, and younger foliated to massive, generally composite plutons, ranging from quartz diorite to granite and syenite. Supracrustal rocks of the greenstone belts include komatiitic, tholeiitic, calc-alkalic, and rare alkalic volcanics with volcanogenic clastic (wacke, conglomerate) and chemical (iron formation, chert) sediments. Most greenstone belts consist of several lensoid, overlapping piles each on the order of 100 km in maximum dimension and approximately 5 to 10 km thick and commonly comprising several volcanic cycles. Some cycles consist of a lower komatiitic -tholeiitic basalt sequence, a middle tholeiitic basalt -andesite sequence, and an upper calc-alkalic dacite-rhyolite-andesite sequence. Other cyles are bimodal tholeiitic basalt-dacite (rhyolite) sequences. Minor alkalic and shoshonitic volcanics and associated alluvial/fluvial sediments are present in some belts where they unconformably overlie older volcanics and synvolcanic plutons. In term of rock types, sequences, and overall configuration, many Superior Province greenstone belts are comparable to modern island arcs.
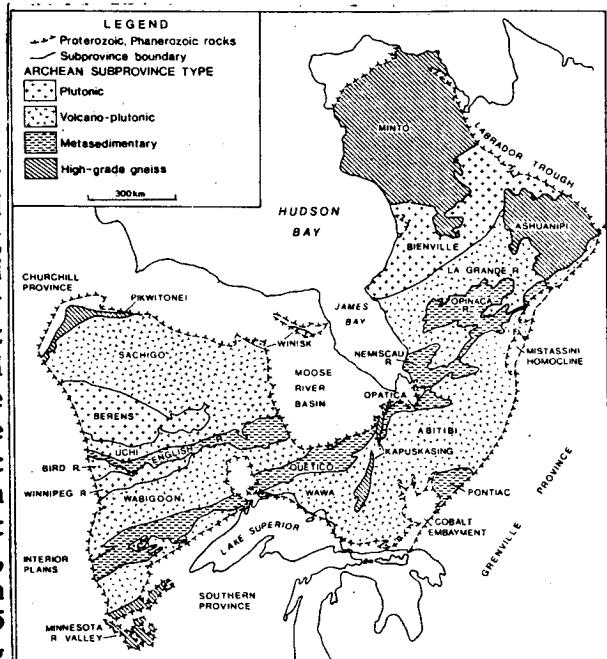
Superior Province greenstone belts typically have upright folds with curved, bifurcating axial surfaces, steep foliations and lineations, and major domal culminations and depressions, the products of polyphase deformation. Some belts, however, display low angle foliations and faults, overturned sequences, and recumbent and downward - facing structures suggestive of thrust-nappe style tectonics (1,9,10).

The enclosing gneissic and plutonic rocks display domal structural patterns, again the product of polyphase deformation involving recumbent folding and diapirism. Metamorphic grade in the greenstone belts is generally subgreenschist to greenschist in the central parts grading outward to low pressure amphibolite facies in belt margins and surrounding plutonic gneisses.

TECTONIC SETTING AND EVOLUTION OF LATE ARCHEAN
K.D. Card

The contacts between greenstone belts and enclosing plutonic rocks, and between the greenstone-rich subprovinces and adjacent plutonic subprovinces are generally either intrusive or tectonic. An unconformity between greenstones and older granitoid rocks has been demonstrated only at Steeprock, Ontario(5) and although younger volcanics and older plutonic rocks are juxtaposed in a number of places, faults, mylonites, or shear zones invariably intervene. Dextral transcurrent faults trending EW and NW and sinistral faults trending NE form subprovince boundaries in part, as do NE and EW trending thrusts. One notable product of this faulting, the Kapuskasing Structural Zone, exposes granulites considered to represent upthrusted lower crust (7,8). Late alkalic volcanic - fluviatile sediment sequences are spatially related to major transcurrent faults and may represent deposition in pull-apart basins formed by alternating periods of transtension and transpression in strike-slip zones.

Interpretation of geophysical data shows changes in depth to the Conrad Discontinuity and to the Moho from one subprovince to another, indicating significant structural relief across their faulted boundaries (4). Greenstone belts of Abitibi and Wabigoon subprovinces generally extend to depths of only 5 to 10 km(3) whereas metasedimentary gneisses of English River Subprovince and plutonic rocks of Winnipeg River Subprovince may extend to depths of 10 to 20 km(4). Juxtaposition of high-pressure granulites of the Kapuskasing zone with low-pressure greenschist-amphibolite facies rocks of Abitibi Subprovince implies structural relief of 15 to 20 km across the boundary thrust (7,8).

LEGEND
Fault, unclassified
Thrust fault
Transcurrent fault
Normal fault
Province boundary
300 km
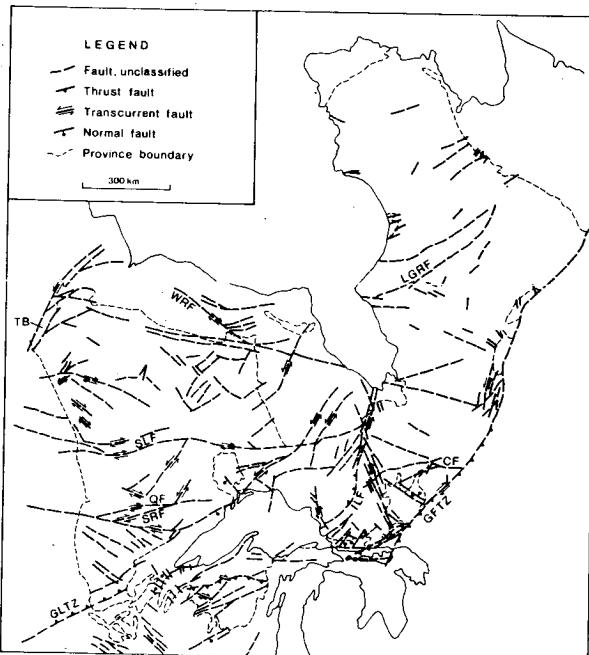
Metasedimentary subprovinces (English River, Quetico, Pontiac etc.) consist mainly of turbidite wacke and pelite metamorphosed at grades ranging from low greenschist at belt margins to upper amphibolite and locally, low-pressure granulite in belt interiors. Anatectic, s-type granitic rocks are prevalent in the migmatitic, high-grade interiors of the metasedimentary belts. Most metasedimentary subprovinces have a linear aspect attributable to transcurrent boundary faults and isoclinal folds with subhorizontal to subvertical axes, late structures superimposed on earlier complex, recumbent folds and dome-basin structures. In areas where contacts between metasedimentary and volcano-plutonic subprovinces are unfaulted, there appear to be rapid facies transitions from sedimentary to dominantly volcanic sequences. Preliminary isotopic age data also indicate that the sedimentary and volcanic sequences of some adjacent subprovinces are broadly coeval.

U-Pb zircon dates demonstrate that volcanic, plutonic, deformational, and metamorphic events of relatively brief duration affected large parts of Superior Province and that there are detectable differences in ages of these events from one area to another(6). In the northwest (Sachigo, Berens, Uchi subprovinces) major volcanism and

accompanying plutonism occurred at about 3.0 to 2.9 Ga, 2.85 to 2.80 Ga, and 2.75 to 2.7 Ga. These voclanic episodes were followed by major deformation, metamorphism, and plutonism about 2.73 to 2.7 Ga. In the south (Wabigoon, Wawa, Abitibi subprovinces) volcanism and plutonism occurred mainly between 2.75 and 2.69 Ga, followed by major deformation, metamorphism, and plutonism at about 2.70 to 2.66 Ga. There is evidence for somewhat younger (2.65 to 2.63 Ga) metamorphic-plutonic events, or of later closure of isotopic systems, in the high-grade rocks of the metasedimentary belts and of the Kapuskasing zone.

In summary, Superior Province consists mainly of Late Archean supracrustal and plutonic rocks with Middle and Early Archean gneisses in the south and possibly in the north. The Late Archean supracrustal sequences are probably mainly of oceanic or marginal oceanic affinity (island arc, marginal basin, submarine plateaus), although continental arc and rift zone settings have also been postulated(2). Abundant plutonic rocks include early synvolcanic intrusions and later synorogenic and post-orogenic intrusions derived in part from the mantle and in part from crustal melting caused by thermal blanketing of newly-thickened continental crust combined with high mantle heat flux. Some pre-greenstone plutonic rocks may represent accreted microcontinents.

The contemporaneity of magmatic and deformational events along the lengths of the belts, coupled with the structural evidence of major compression and transcurrent faulting, is consistent with a subduction-dominated tectonic regime for assembly of the Superior Province orogen. Successive lateral and vertical accretion of volcanic arcs and related sedimentary accumulations, accompanied and followed by voluminous plutonism, resulted in multi-stage crustal thickening and stabilization of the Superior craton prior to emplacement of mafic dyke swarms and Early Proterozoic marginal rifting.

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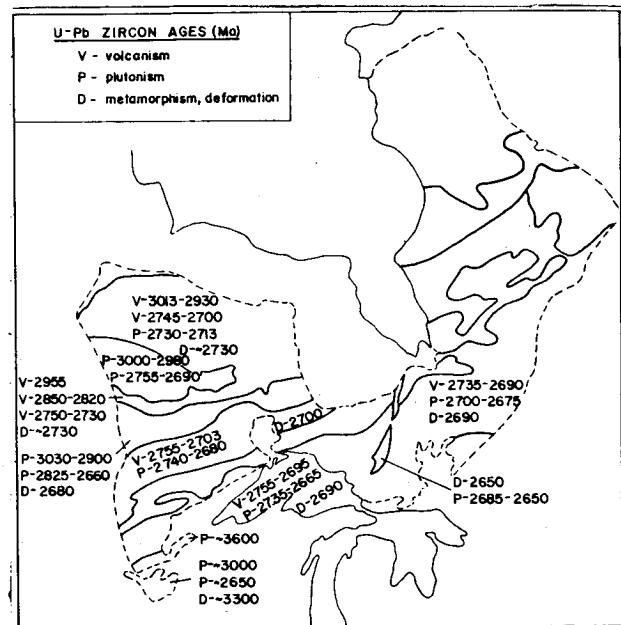
HIGH PRECISION U-PB GEOCHRONOLOGY AND IMPLICATIONS FOR THE TECTONIC EVOLUTION OF THE SUPERIOR PROVINCE; D.W. Davis, F. Corfu, and T.E. Krogh, Jack Satterly Geochronology Laboratory, Royal Ontario Museum, 100 Queen's Park, Toronto, Ontario. M5S 2C6

The underlying mechanisms of Archean tectonics and the degree to which modern plate tectonic models are applicable early in earth's history continue to be a subject of considerable debate. A precise knowledge of the timing of geological events is of the utmost importance in studying this problem. The high precision U-Pb method has been applied in recent years to rock units in many areas of the Superior Province. Most of these data have precisions of about ± 2-3 Ma. The resulting detailed chronologies of local igneous development and the regional age relationships furnish tight constraints on any Archean tectonic model.

Superior province terrains can be classified into 3 types:
1) low grade areas dominated by meta-volcanic rocks (greenstone belts).
2) high grade, largely metaplutonic areas with abundant orthogneiss and foliated to massive I-type granitoid bodies.
3) high grade areas with abundant metasediments, paragneiss and S-type plutons.

Most of the U-Pb age determinations have been done on type 1 terrains with very few having been done in type 3 terrains.

A compilation of over 120 ages indicates that the major part of igneous activity took place in the period 2760-2670 Ma, known as the Kenoran event. This event was ubiquitous throughout the Superior Province.

There is, however, abundant evidence for the widespread occurrence of pre-Kenoran volcanoes and sialic crust, especially north of the Wabigoon-English River subprovince boundary. In the Uchi and Sachigo subprovinces there are volcanic periods about 3000-2900 Ma (2,3) and 2850-2800 Ma (2,3) in age which underlie the Kenoran sequence. The Kenoran rocks are in part disconformable on the older sequences. The general absence of angular unconformities along with other evidence such as the presence of mature sandstones in the Sachigo subprovince (4), implies an extended period of crustal stability preceding the Kenoran event. Tonalites 2950-3200 Ma in age are found in the Favourable Lake (5), North Spirit Lake (3) and Winnipeg River Belts (6,7,8), suggesting a pre-Kenoran crust-forming event at about 3000 Ma. Evidence for the existence of an extensive pre-Kenoran continent is especially strong in the Winnipeg River Belt, a type 2 terrain. Recent data obtained from a type 2 terrain in the Wabigoon subprovince also indicates 3000 Ma volcanic and plutonic sequences (9). This indicates that type 2 terrains in many cases include pre-Kenoran crust, and that pre-Kenoran crustal material may be locally present in type 2 areas throughout the Superior Province.

The earliest Kenoran magmatism consisted of eruption of tholeiitic basalt platforms. These are difficult to date but in some areas pre-date 2750 Ma (10). Intermediate-felsic calc-alkaline volcanism occupied a time span from about 2750-2700 Ma and led to the construction of large composite volcanoes. The transition to calc-alkaline volcanism was associated with the emplacement of layered basic intrusions and contemporaneous tonalite-granodiorite plutons, without major deformation. This was followed by the intrusion of high alumina trondhjemite-granodiorite plutons, in some cases accompanied by later calc-alkaline volcanism. This resulted in the development of large intravolcanic batholiths (11). Significant regional

deformation began relatively late, at about 2700 Ma over much of the Superior Province. In some areas, such as the Wabigoon greenstone belt, it significantly post-dated the bulk of calc-alkaline intrusive and volcanic activity (12).

Some greenstone belts underwent at least two periods of deformation. A D1 event affected the calc-alkaline sequences and pre-dated sedimentation and eruption of alkaline volcanic rocks (e.g. Tamiskaming sequence). The Tamiskaming-type sequences were then affected by a later deformational (D2) event. The ages of the late sequences and the D2 event are bracketed between 2695 and 2685 Ma in the southern Wabigoon and Shebandowan sub-provinces (13), but may have been 15-20 Ma earlier in the Oxford Lake belt in the northern Superior Province.

The causes of regional deformation are unclear. It may have been partly the result of diapiric remobilization of the intravolcanic batholiths (e.g. Wabigoon greenstone belt), accompanied by regional compression, perhaps due to the intrusion of marginal late granitoid plutons (e.g. Batchawana belt) (14). The presence of nappe structures in some areas such as the Winnipeg River belt (15) and the southern Wabigoon subprovince (16) further complicates the tectonic picture. The final expression of strain was the establishment of large strike-slip faults, which often separate type 1 from type 2 and type 3 terrains.

The deformational event was accompanied by intrusion of late tectonic plutons, most of which have ages in the range 2700-2670 Ma (17). This resulted in cratonization and brought the Kenoran event to an end. Locally, single volcanic centers passed through this cycle from initial volcanism to terminal deformation in time spans as short as 30 Ma (18).

Although there is some indication of a secular younging in the peak of Kenoran igneous activity in a N-S direction, the broad simultaneity and short time spans of crustal events argue against any simple model for growth of the Superior Province by accretion of island arcs (19). Furthermore, there is a strong vertical control on magmatic and metamorphic ages. The oldest Kenoran plutons occur high in the crust while the youngest plutons and metamorphic ages are found at deeper crustal levels in more uplifted and eroded terrains such as the Berens River subprovince, parts of the Winnipeg River belt and the Kapuskasing structural zone (20).

Despite considerable work on the felsic units in type 1 greenstone terrains, there is almost no evidence of inherited zircon components derived from significantly older sialic material. The intravolcanic granitoid rocks and thick felsic volcanic sequences were largely derived by differentiation processes from mafic precursors within the period of Kenoran activity (11). However, greenstone belts evidently did develop adjacent to older sialic blocks. Evidence for this, found in the Wabigoon greenstone belt, includes pre-Kenoran granitoid clasts in a conglomerate marginal to the belt and the existence of marginal unconformities between 3000 Ma tonalite in the Winnipeg River belt and Kenoran volcanic and plutonic sequences (7). Abundant mafic dykes intrude the older units below these unconformities and indicate a tensional stress regime.

The bulk of the evidence presently available argues for a model in which greenstone belts were initiated by rifting of older sialic crust and the formation of narrow ocean basins. The fault controlled nature of many subprovince boundaries as well as the fact that volcanism was at times nearly coeval throughout the Superior Province suggests that rifting may have been concentrated along major early lithospheric breaks.

Evidence for subduction in late Archean tectonic processes is missing.
The absence of an effective subduction mechanism would have inhibited ocean
spreading. If the intracratonic rifts were not able to open into wide ocean
basins they would have been reworked in place, undergoing dominantly vertical
tectonic processes. Continued mantle-derived mafic magmatism may have led to
thickening and differentiation of the crust to produce the large amounts of
calc-alkaline material now present in type 1 terrains.

Any model for tectonic development can only be tentative and subject to
the constraints of a constantly expanding data set. Some of the major
questions remaining for geochronology are the extent in time and space of
pre-Kenoran material and its deformational history and the origin and
basement of the metasedimentary belts. These questions can only be resolved
by much more extensive work in type 2 and type 3 terrains.

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CRUSTAL STRUCTURE OF THE ARCHEAN GRANITE-GREENSTONE TERRANE IN THE
NORTHERN PORTION OF THE KAAPVAAL CRATON; J.H. de Beer (NPRL, CSIR, P.O. Box
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Recent investigations of the electrical resistivity, gravity and aeromag-
netic signatures of the various granite-greenstone units in the northern por-
tion of the Kaapvaal Craton have revealed three features of significance:
1) The Archean greenstone belts are shallow features, rarely exceeding
   5 km in depth;
2) The high resistivity upper crustal layer typical of the lower-grade
   granite-greenstone terranes is absent in the granulite facies terrane,
   and
3) The aeromagnetic lineation patterns allow the granite-greenstone ter-
   rane to be subdivided into geologically recognizable tectono-metamor-
   phic domains on the basis of lineation frequency and direction.

In the Pietersburg, Sutherland and Murchison Greenstone Belts, geoelect-
rical investigations showed that the greenstone lithologies have a lower re-
sistivity than the surrounding granitoid terranes. Positive gravity anomalies
over the greenstone belts are related to more dense metamorphosed ultramafic
and mafic rocks in the Belts compared to surrounding granitoid rocks. Numerical
modelling of the geophysical data indicates that the greenstone belts are asym-
metrical structures, being thicker along the southeastern flanks. The belts
are underlain by high resistivity, low density granitoid rocks of which two
types are distinguished by their average densities:
1) a lower density series (density = 2 600 kg/m) corresponding to 2650 Ma
   granodioritic plutons and
2) a higher density series (density = 2 670 kg/m) comprising the older
   tonalitic and trondhjemitic gneisses.
The younger series is particularly well developed along the southern margins of
the greenstone belts and occurs locally along the northern margins. Primary
layering and tectonic fabrics within the greenstone lithologies are subvertical.
Thicknesses measured across layering exceed the depth of the belts, suggesting
no simple rotation of the greenstone lithologies but instead a truncation at
shallow depths of structurally repeated (folded and imbricated) greenstone
belts. These truncations may be major recumbent deformation zones, recumbent
syntectonic granitoid intrusions or a late intrusive contact.

Deep resistivity soundings indicate significant changes in the regional
structure of the crust in the northern portion of the Kaapvaal Craton corres-
ponding to changes in metamorphic grade and tectonic style. In the low-grade
granite-greenstone terrane, the upper 10 km or less of the crust is sharacter-
ized by high-resistivity rocks (approximately 100 000 ohm metre) overlying a
more conductive layer (approximately 5 000 ohm metre) to a depth of about 35
km. Below this depth, possible mantle rocks with a resistivity of about 50
ohm metre occur. Where the granulite facies of the Southern Marginal Zone of
the Limpopo Belt occur, the approximately 100 000 ohm metre layer is absent and
rocks with a resistivity of about 5 000 ohm metre extend to a maximum depth of
about 35 km where they overlie possible mantle rocks. Work in progress indi-
cates that in the vicinity of the orthoamphibole isograd marking the southern
boundary of the Southern Marginal Zone, the high-resistivity upper crustal
layer characterizing the low-grade terrane to the south dips northward under-
neath the moderately resistive high-grade rocks with lower crustal signatures.

CRUSTAL STRUCTURE IN THE NORTHERN KAAPVAAL CRATON

J.H. de Beer et al.

This relationship is consistent with a crustal model in which the lower crustal rocks of the high-grade terrane have over-ridden the upper crustal rocks of the low-grade terrane. Some of the gravity models calculated for the Sutherland Greenstone Belt are also in agreement with such a tectonic model. At this stage of the research, it is not clear how the limited seismic data and iso-static gravity data relate to such a structural model.

The aeromagnetic lineation pattern in the northern portion of the Kaap-vaal Craton can be divided into distinct domains on the basis of the lineation frequency and direction. Although these magnetic anomalies are due to mafic and ultramafic dykes, they reflect an inherent fabric in the crust. The do-main boundaries correspond to known tectonic and/or metamorphic transitions. One such boundary is the orthoamphibole rehydration isograd that marks the transition between the granulite facies terrane of the Southern Marginal Zone and the lower-grade rocks to the south. It is clear that the lineation pat-tern does not reflect the different lithological units in the area and the Sutherland and Pietersburg Greenstone Belts are, for example, not reflected in the aeromagnetic lineation pattern. This suggests that these greenstone belts are internal components of certain domains and do not mark domain sutures.

FELSIC VOLCANICLASTIC ROCKS IN THE 3.3 TO 3.5 Ga WARRAWOONA GROUP, PILBARA BLOCK, WESTERN AUSTRALIA: DEPOSITIONAL SETTING AND CRUSTAL EVOLUTION; DiMarco, M.J., and Lowe, D.R., Dept. of Geology, Louisiana State University, Baton Rouge, LA 70802 USA

The 3.3 to 3.5 Ga Coongan Formation (1), a newly identified, predominantly volcaniclastic unit in the eastern Pilbara Block, Western Australia, records sedimentation patterns associated with the development of felsic volcanism and provides important information on early terrestrial crustal evolution. The Coongan Formation is divided into three members: The Duffer and Panorama Members and the Strelley Pool Chert. The Duffer Member, formerly known as the Duffer Formation (2,3), is the oldest and volumetrically greatest member of the Coongan Formation and consists of felsic volcaniclastic rudite and felsic lava with lesser amounts of turbiditic tuff, tuff-breccia, and chert. The Duffer Member is overlain by or interfingers in its upper part with the Panorama Member, formerly known as the Panorama Formation (2,3), consisting predominantly of silicified volcaniclastic arenite with lesser amounts of felsic tuff, silicified lutite, chert, felsic lava, chert-clast conglomerate, volcaniclastic rudite, and barite. The Duffer and Panorama Members are overlain unconformably (1) by the Strelley Pool Chert (4), which represents silicified orthochemical and biogenic sediments.

Rocks of the Duffer Member represent volcaniclastic sedimentation on coarse debris-aprons that flanked felsic volcanic centers. These aprons are made up primarily of debris-flow deposits with lesser amounts of intercalated turbidites. Most of the debris-aprons were subaqueous features, but, locally, they are capped by sequences of alluvial conglomerate, indicating that they shoaled upward with time. Predominantly arenaceous strata of the Panorama Member represent reworked, and to a lesser extent, direct-deposited pyroclastic debris that was dispersed to a spectrum of shallow-water and subaerial environments. Tephra production during Panorama times was probably related to the generation of increasingly more evolved magmas during the later stages of felsic volcanism (5,6). Evidence of several features typical in continental felsic volcanic provinces, including caldera complexes and thick sequences of ash-flow tuffs, was not seen in the Duffer and Panorama Members. After a period of uplift and erosion, orthochemical and biogenic sediments, represented by the Strelley Pool Chert, were deposited on a broad, post-volcanic platform.

Clastic rocks of the Coongan Formation indicate derivation from a largely felsic volcanic provenance. Clasts in rudite of the Duffer Member consist predominantly of felsic volcanic rock. Locally, a few rudite units contain sparse clasts of chert. Framework modes of arenites of the Panorama Member consist mainly of altered felsic volcanic rock fragments with subordinant amounts of volcanic quartz, chert, and pseudomorphs after feldspar and minor amounts of altered mafic volcanic rock fragments and pseudomorphs after mica and amphibole. No evidence for detrital contributions from granitic sources was found in the Duffer and Panorama Members. In addition to the predominant felsic volcanic provenance, minor contibutions were received from mafic volcanic and local sedimentary sources.

Paleocurrent evidence, mainly from cross-bedded arenites and imbricate conglomerates in the Duffer and Panorama Members, indicate sediment dispersal from several felsic source areas close to the present location of granitic plutons that crop out between greenstone belts in the eastern Pilbara. These data support earlier proposals, based primarily on geochemical and

VOLCANICLASTIC ROCKS IN THE WARRAWOONA GROUP
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and isotopic data (6,7), that some of the granitic plutonic rocks and felsic volcanic rocks in the eastern Pilbara are coeval. These felsic centers were apparently randomly distributed, unlike those in many modern volcanic arcs.

Despite earlier suggestions of a primordial, pre-greenstone granitic crust (2), no direct evidence for rocks older than the Warrawoona Group has been reported. The lack of evidence for detrital contributions from plutonic sources in the Duffer and Panorama Members indicates that the granitic rocks were not yet exposed in Coongan times. The petrographic and paleocurrent evidence presented here coupled with earlier geochemical and isotopic work (6,7) suggest that the felsic volcaniclastic and volcanic rocks of the Coongan Formation represent the volcanic component of evolving granitic plutonism and the earliest manifestation of cratonization in the eastern Pilbara Block. Granitic intrusion culminated with tectonism and exposure of the plutons during Gorge Creek times, when sediments containing granitic detritus were deposited (2,3,7).

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EXTENSIONAL TECTONICS DURING THE IGNEOUS EMPLACEMENT OF THE MAFIC-ULTRAMAFIC ROCKS OF THE BARBERTON GREENSTONE BELT. M.J. de Wit, Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058 and BPI Geophysics, University of the Witwatersrand, Johannesburg, South Africa.

The simatic rocks (Onverwacht Group) of the Barberton greenstone belt, which occur in at least 3 regional thrust nappes, are part of the Jamestown ophiolite complex[1]. This ophiolite, together with it's thick sedimentary cover (Fig Tree and Moodies Groups) occupies a complex thrust belt. Field studies have identified two types of early faults which are entirely confined to the simatic rocks and are deformed by the later thrusts and associated folds. The first type of fault (F1a) is regional and always occurs in the simatic rocks along and parallel to the lower contacts of the ophiolite-related cherts (Middle Marker and equivalent layers; for their distribution see Fig. 1, de Wit et al., this volume). These faults zones have previously been referred to both as flaser-banded gneisses[2] and as weathering horizons[3]. (F1a) zones consist of anastomosing, cross-cutting and folded extension veins which have internal cross-fibrous growth textures. Vein filling minerals are predominantly calcite, less often quartz. The veins are separated by schistose to proto-mylonitic folia of fuchsite, chlorite, sericite and serpentines (Fig. 1). In general the zones range between 1-30m in thickness. The veins formed by a succession of dilation-diffusion increments[4,5] and subsequently deformed during simple shear to form banded gneisses (Fig. 1; in this poster presentation, polished slabs of these rocks will be displayed). The simatic host rocks close to (F1a) zones, are ubiquitously brecciated and extensively altered (carbonatized and/or silicified) as documented by the major elements, stable and radiogenic isotope compositions (REE are relatively stable). This alteration is related to a extensive hydrothermal-fluid/rock interaction. It has been postulated that the dilatancy-anisotropy of the fault zones was related to a hydraulic fracturing-gliding mechanism in a geothermal environment[6]. Episodic decrease of fluid overpressure due to movement in these zones would cause boiling, calcite precipitation and crack-sealing with a concomittant resistance to movement of the cherty cap-rock[6]. Displacements along these zones are difficult to estimate, but may be in the order of $1-10^2$ km. The structures indicate that the faults formed close to horizontal, during extensional shear and were therefore low angle normal faults. In many areas, both the faults and their overlying cherts, are cut by subvertical simatic intrusions of the Onverwacht Group (Fig. 2). Thus (F1a) zones overlap in age with the formation of the ophiolite complex. The second type of faults (F1b) are vertical brittle-ductile shear zones, which crosscut the complex at variable angles and cannot always be traced from plutonic to overlying extrusive (pillowed) simatic rocks. (F1b) zones are therefore also apparently of penecontemporaneous origin with the intrusive-extrusive igneous processes (Fig. 2). Thus (F1b) zones may either represent transform fault-type activity or represent root zones (steepened extensions) of (F1a) zones. Both fault types indicate extensive deformation in the rocks of the greenstone belt prior to compressional overthrust tectonics, and at least (F1a) implies regional extensional tectonics and probably block rotation during the formation of the ophiolites.

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de Wit, M. J.
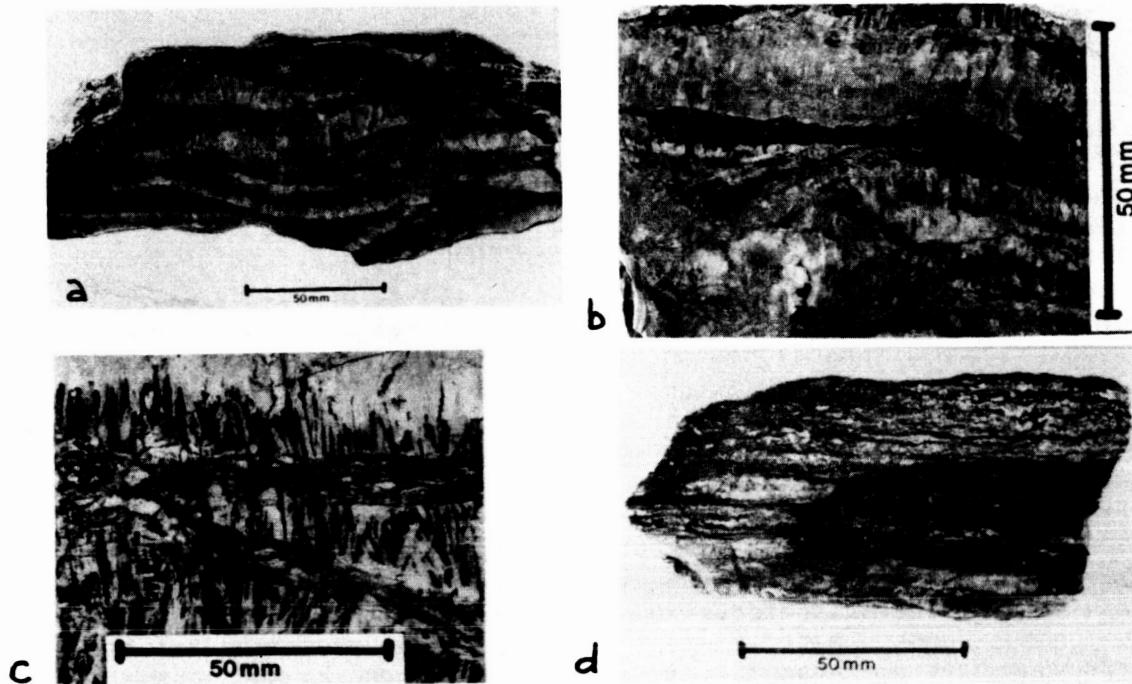
FIG1
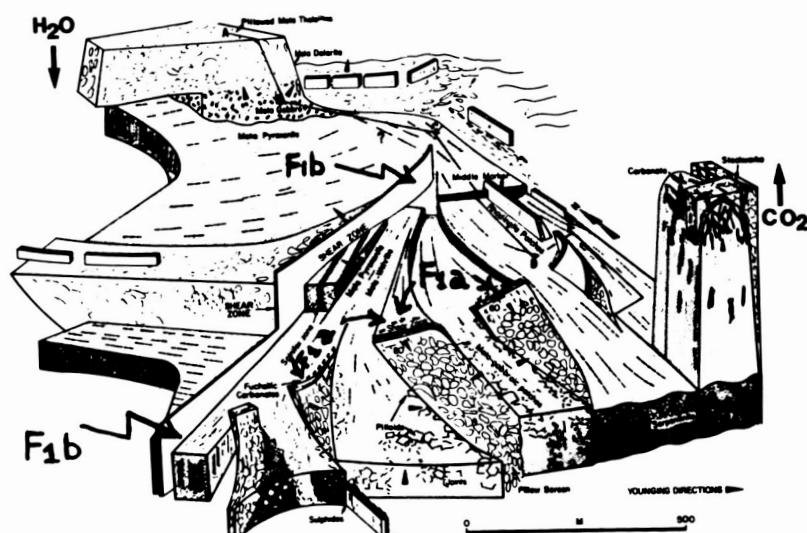
FIG. 2

Figure Captions: Fig. 1 (a) Anastomosing/crosscutting carbonate extension veins (pale-grey) with thin schistose folia (dark grey). Sections up to 30 meters thick entirely composed of this rock-type constitute f laser-banded tectonites[2]. (b) as in (a), showing the cross-fibrous carbonate growth textures in the veins. Different shades of grey are due to variations in concentration of inclusion bands and trails[4]. (c) Internal brecciation and shearing of cross-fibres (vertical) yielding (subhorizontal) protomylonites. (d) gneissose-mylonitic fabric following shearing and flattening of extension veins. (2) Block diagram of area near the Onverwacht bend (see Fig. 1, de Wit this volume for location - as outlined by the box marked Fig. 2) showing the disposition of both (F1a and F1b) fault zones. Note how vertical metawehrlite intrusions (grey) have cross-cut and incorporated screens of middle marker-like cherts underlain by (F1a) gneissose tectonites (shear zones).

A MID-ARCHEAN OPHIOLITE COMPLEX, BARBERTON MOUNTAIN LAND; M.J. de Wit, Lunar and Planetary Institute, 3303 NASA Rd. One, Houston, TX 77058 and BPI Geophysics, University of the Witwatersrand. Roger Hart, School of Oceangraphy OSU, Corvallis, Oregon 97331. Rodger Hart, SC Nuclear Sciences, University of the Witwatersrand, Johannesburg, South Africa.

New field observations and structurally restored geologic sections through the southern part of 3.5-3.6 Ga Barberton greenstone belt (Fig. 1) show that it's mafic to ultramafic rocks form a pseudostratigraphy comparable to that of Phanerozoic ophiolites; we refer to this ancient ophiolite as the Jamestown ophiolite complex[1]. It consists of an (in part sheeted, Fig. 2) intrusive-extrusive mafic-ultramafic section, underlain by a high-temperature tectono-metamorphic residual peridotitic base, and is capped by a chert-shale sequence which it locally intrudes. Geochemical data support an ophiolitic comparison (Fig. 3). Fractionation of high temperature melting PGE's (> $2500^{\circ}C$) in the residual rocks suggest a lower mantle origin for the precursors this crust[2]. An oceanic rather than arc-related crustal section can be inferred from the absence of contemporaneous andesites. This ancient simatic crust was thin (<3 km), contains a large ultramafic component ($\approx 25\%$), is pervasively hydrated (> 95%) with H_2O contents ranging between 1-15% and consequently has a low density (≈ 2.67 g/cm^3)[3].

The entire simatic section has also been chemically altered during its formation by hydrothermal interaction with the Archaean hydrosphere (Fig. 4). Only an igneous "ghost" major element geochemistry is preserved. This regionally open-system metasomatism may have increased the MgO content of the igneous rocks by as much as 15%. The most primitive parent liquids, from which the extrusive sequence evolved, may have been "picritic" in character. Rocks with a komatiitic chemistry may have been derived during crystal accumulation from picrite-crystal mushes (predominantly olivine-clinopyroxene) and/or by metasomatism during one or more subsequent episodes of hydration-dehydration (Fig. 5).

The Jamestown ophiolite complex provides the oldest record with evidence for the formation of oceanic lithosphere at constructive tectonic boundaries. Our observations are in agreement with models predicting higher oceanic Archean heat flux per unit ridge length than today, associated with deep mantle diapiric upflow. Because of its low density, this ophiolite resisted subduction during subsequent tectonism; it was obducted to form part of a thrust complex[3].

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A MID ARCHEAN OPHIOLITE COMPLEX

de Wit, M. J. et. al.

FIG 1

FIG2

a

FIG 3

b

A MID ARCHEAN OPHIOLITE COMPLEX

de Wit, M. J. et. al.
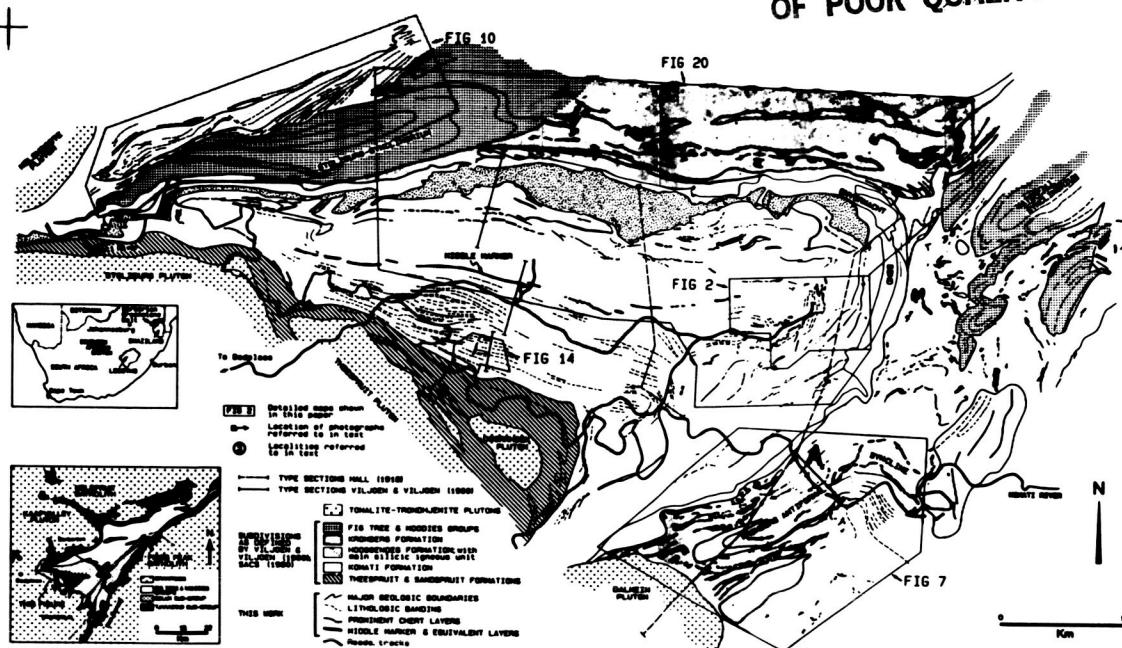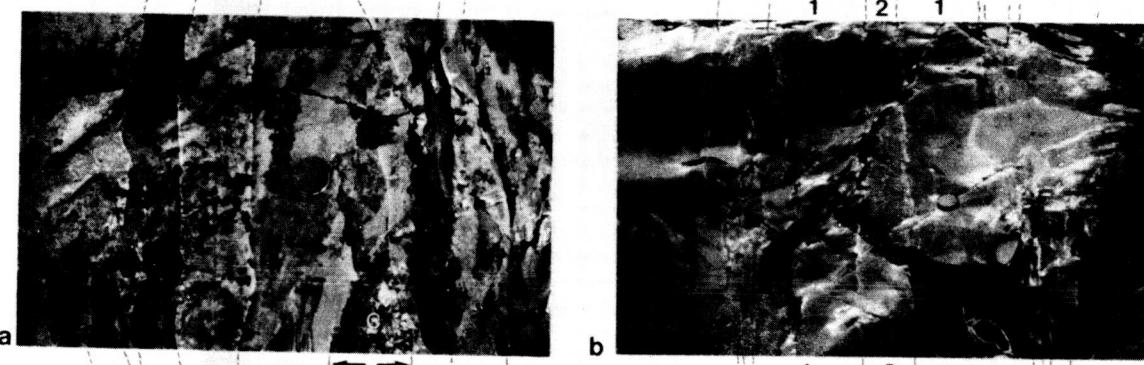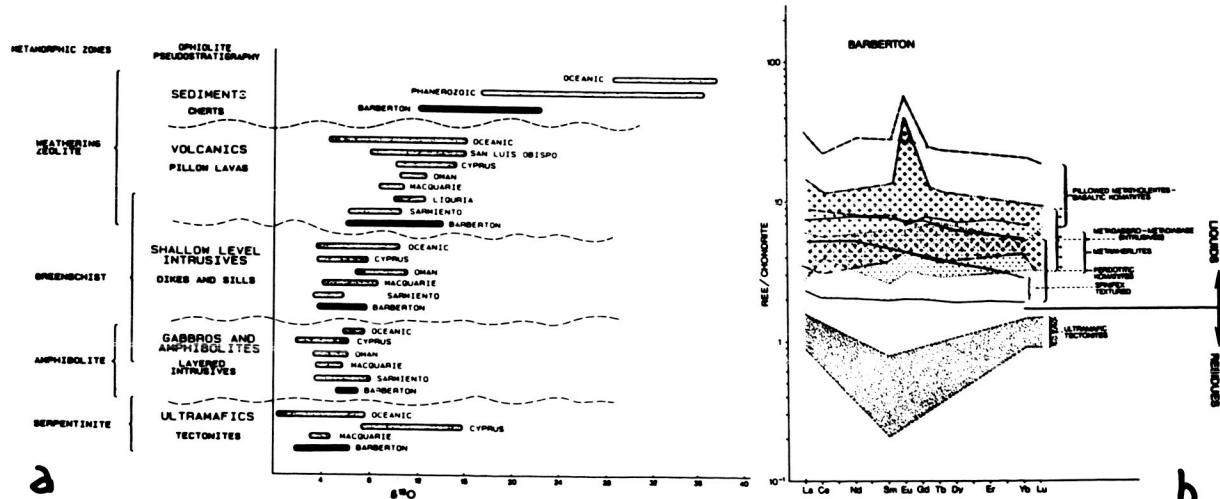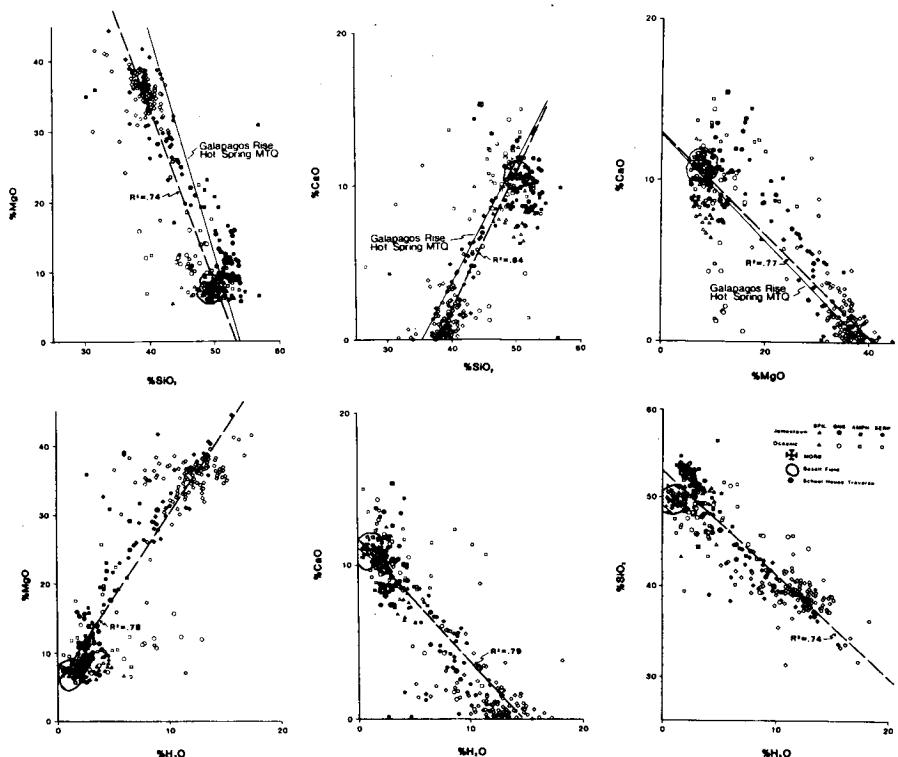
FIG.4
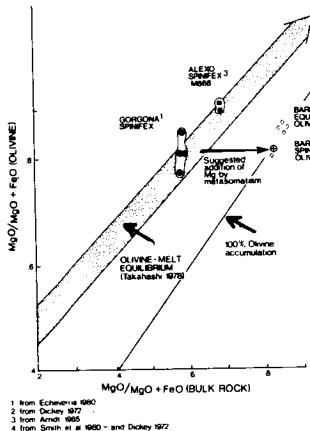
FIG.5

Figure Captions: (1) Simplified geological map of an area in the southern part of the Barberton greenstone belt studied between 1978 and 1985. (2) Vertical sheeted intrusives with pale chilled margins from a 30 meter river outcrop (exposed during 1984 draught) at locality A, Fig. 1. Note the remnant chert xenolith (a; arrow). (b) clearly depicts the splitting in two of an earlier intrusion (1) by a later one (2). (3) Representation of $\delta^{18}O$ (a) of the Barberton rocks (black) plotted in their restored pseudo-stratigraphic sequence compared to Phanerozoic ophiolites and oceanic crust (open symbols) (b) REE data from Barberton; this plot compares favourably with Phanerozoic ophiolites and oceanic crust. (4) Binary correlation plots of MgO, CaO, SiO, and H_2O for rocks of oceanic crust (open symbols) and from the Jamestown ophiolite complex (closed symbols). These plots illustrate the close correlation between the major oxides concentrations and the degree of hydration in these environments. For comparison, the slopes of the chemical flux in the Galapages hot spring fluids are also shown. (5) This figure shows that the bulk rock MgO/MgO + Feo ratio of Barberton Komatiites are enriched in MgO over that of the original melt. The enrichment may be the result of either crystal accumulation or magnesium metasomatism during hydrothermal alteration; there is textural evidence that both mechanism were important. At any rate the plot clearly shows that the MgO composition of the silicate liquids which formed the Barberton Komatiite was between that of Gorgona Island (15-22% MgO) and Alexo (28% MgO), and may have been of picritic composition. All diagrams from reference 1.

FELSIC IGNEOUS ROCKS WITHIN THE BARBERTON GREENSTONE BELT: HIGH CRUSTAL LEVEL EQUIVALENTS OF THE SURROUNDING TONALITE-TRONDHJEMITE TERRAIN, EMPLACED DURING THRUSTING.

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Felsic rocks within the 3530 + 50 myrs[1] simatic rocks of the Onverwacht Group of the Barberton greenstone belt have traditionally been mapped as recurring volcanic units within a continuous stratigraphic succession. In the past, these felsic units have been interpreted to be part of several mafic to felsic volcanic cycles within this sequence. Some of these silicic layers have been shown to be silicified simatic rocks[2,3]. Our field data (Fig. 1) indicates that the genuine felsic igneous rocks are predominantly shallow level intrusives and subsurface felsic domes associated with only minor volcanics and volcanoclastics. A 3.360 + 1 myrs (U-Pb, zircon)[4] age from the main felsic intrusion indicates that its emplacement post-dated the simatic rocks of this greenstone belt between 120-220 myrs. Our geochemical results also show that the felsic igneous rocks are not directly related to the mafic-ultramafic rocks of the Onverwacht Group. On the contrary the major trace and REE data (Fig. 2) all indicate that these felsic units are high-level equivalents of the widespread, and time-equivalent, trondhjemite-tonalite plutons which either intrude the lower parts of the greenstone belt, or with which they are in tectonic contact.
Structural and stratigraphic analysis indicates that the felsic intrusions were emplaced along thrusts during sedimentation and a prolonged period of horizontal compressional stress exerted on the greenstone belt (Fig. 3). Thus, integrated, the data suggest that the simatic rocks of the Barberton greenstone belt were thrust across an actively stoping plutonic environment and that the greenstone belt is at least partly allochthonous (Fig. 4).

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FELSIC IGNEOUS ROCKS EMPLACED DURING THRUSTING
de Wit, M. J. and Wilson, A. H.

FIG.1

a

b
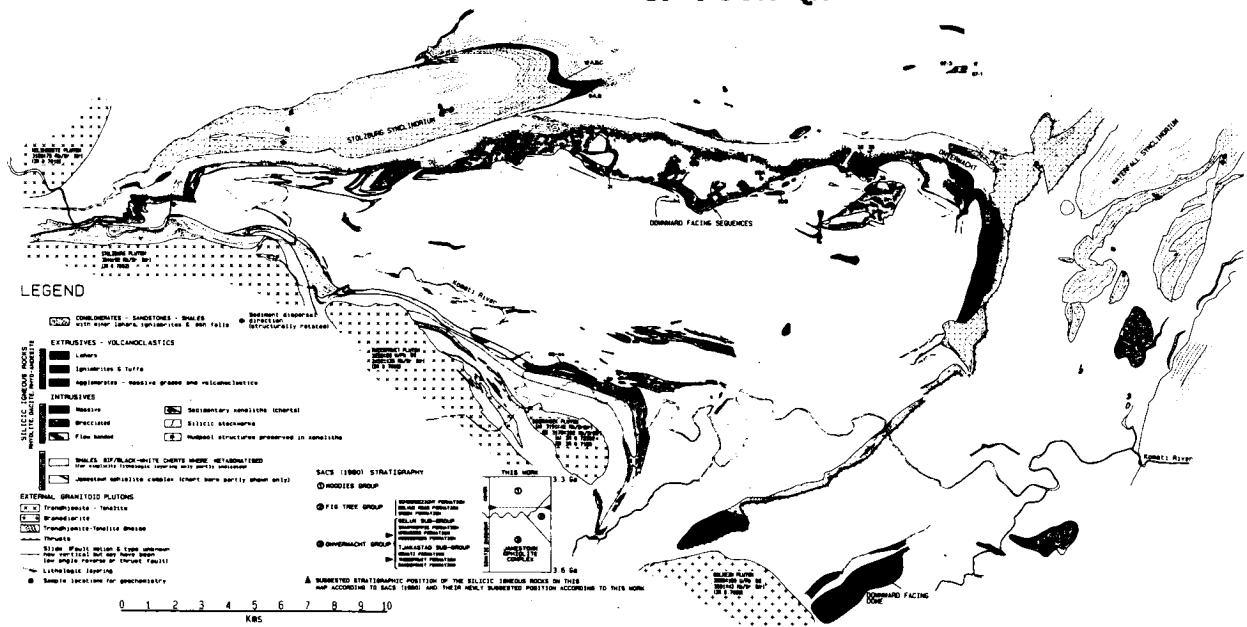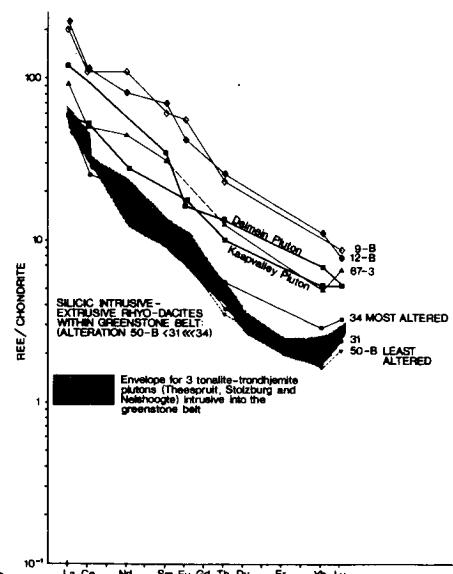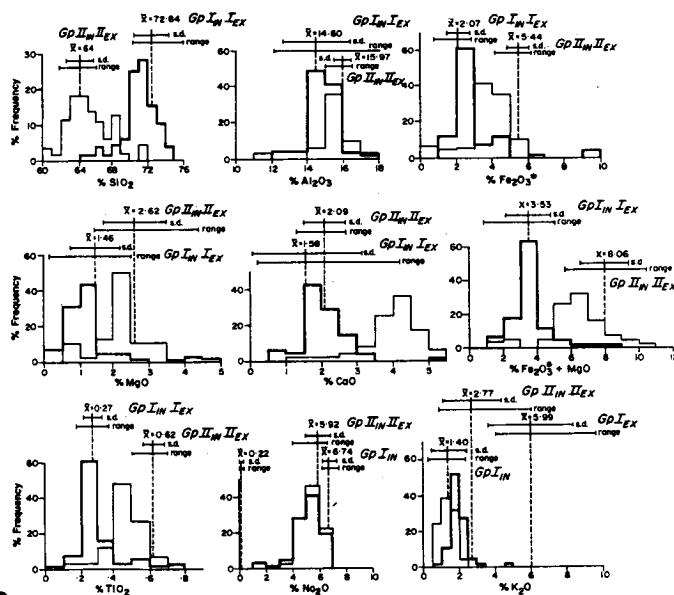
FIG.2

Figure Captions: (1) Simplified geological map of the southern part of the
Barberton greenstone belt, showing location of main silicic (felsic) rocks. (2)
(a) Statistical analysis of major element data from the felsic igneous rocks
within the study area, compared to those of the surrounding tonalite and
trondhjemites. The felsic igneous rocks are clearly divided into two groups (I
and II) in which both extrusive (ex) and intrusive (in) samples are represented.
The two groups are geochemically similar to the trondhjemites (thin frequency
boxes) and tonalites (bold frequency boxes) (b) Chondrite normalized REE
patterns of intrusive and extrusive representatives of both groups of felsic
igneous rocks from within the greenstone belt, compared to the granitoid plutons
surrounding the greenstone belt.

FIG.3

GREENSTONE BELT

GRANITOID/FELSIC ROCKS

FOLIATION

TECTONIC TRANSPORT DIRECTION (D2)

PLUTONS
I STENTOR
II NELSPRUIT
III KAAP VALLEY
IV NELSHOOGTE
V STOLZBERG
VI THEESPRUIT
VII DALMEIN

FIG 4

TONALITE-TRONDHJEMITE MELTS FROM TECTONICALLY BURIED GREENSTONE ALLOCHTHONS
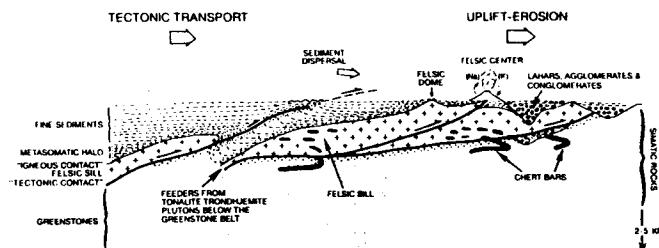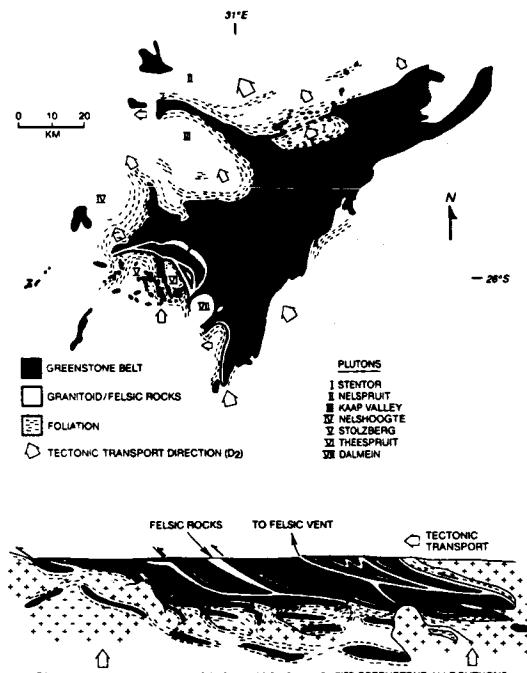
(3) Schematic representation of the tectonic-intrusive emplacement of the felsic igneous rocks as composite sills close to the interface between the mafic-ultramafic (simatic) rocks of the Onverwacht rocks (diagonal lines) and the overlying Fig Tree-like silts-shales. Note how the lower contacts of the sills are predominantly tectonic (thrusts) whilst the upper contacts are predominantly preserved igneous contacts. (4) Plan and section of the Barberton greenstone belt (black) and the surrounding granitoid terrain (white). The map shows the generalized D_2 tectonic transport directions, the felsic igneous rocks internal to the greenstone belt, and the gneissose fabric in the surrounding tonalite-trondhjemite plutons. Note that large scale stoping of the greenstone belt by the surrounding and intruding granitoids is suggested by the outcrop pattern of the felsic igneous rocks (eg. compare this pattern to the shape and outline of the Stentor pluton). The section schematically shows the lower parts (3-5 km) of the greenstone belt thrust over the granitoid terrain whilst the latter syntectonically intrudes and engulves the greenstone belt: this process is thought to have formed recumbent-like mantle-gneiss folds (probably sheath-like in 3-dimentions). Regional disruption and stoping of the greenstone belt occurs during intrusion of Na-rich felsic phases from the plutons of the granitoid terrain into the greenstone belts, along thrusts generated during the tectonic emplacement of the entire greenstone belt. The section represents a restoration prior to subsequent horizontal flattening which later deformed and rotated the rock units and their contacts into a pseudo-synformal structure. All diagrams from de Wit, Wilson and Armstrong (1985 under review).

HEAT FLOW AND HEAT GENERATION IN GREENSTONE BELTS; Malcolm J. Drury, Earth Physics Branch, Energy, Mines and Resources Canada, 1 Observatory Crescent, Ottawa, Ontario, K1A 0Y3 Canada

Heat flow has been measured in Precambrian shields in both greenstone belts and crystalline terrains. Values are generally low, reflecting the great age and tectonic stability of the shields; they range typically between 30 and 50 mW/m², although extreme values of 18 and 79 mW/m² have been reported (1,2). For large areas of the earth's surface that are presumed to have been subjected to a common thermotectonic event, plots of heat flow against heat generation appear to be linear (3,4), although there may be considerable scatter in the data. The relationship is expressed as:

$$Q = Q_0 + D A_0 \qquad [1]$$

in which Q is the observed heat flow, A_0 is the measured heat generation at the surface, Q_0 is the "reduced" heat flow from the lower crust and mantle, and D, which has the dimension of length, represents a scale depth for the distribution of radiogenic elements. Most authors have not used data from greenstone belts in attempting to define the relationship within shields, considering them unrepresentative and preferring to use data from relatively homogeneous crystalline rocks, e.g. (5).

The heat generated by radioactive decay is expected to be less in basic than in acidic rocks because of their different chemistry. Hence we would expect heat flow in greenstone belts to be lower than that in adjoining crystalline areas if the greenstones are thick, but to be similar if the belts are merely superficial. Table 1 is a compilation of data from seven Precambrian shields. Only those data specifically identified as being from greenstone belts, or those for which geological descriptions are unambiguous, are used in column 2. There is the possiblity that some of the data identified as being from crystalline areas are in fact from greenstone belts.

Table 1. Compilation of heat flow data for Precambrian shields, listed according to geological setting. The ratio in column 4 is that of the mean heat flow in the greenstone belts to that in crystalline areas of the shield.

Shield	Mean and 1 s.d. heat flow (mW/m²)			
	All sites	Greenstones	Crystalline	Ratio
Canadian*[a]	42±8 (22)	39±5 (8)	43±10 (14)	0.91
Canadian*[b]	43±10 (10)	40±9 (6)	48±11 (4)	0.83
Baltic[c]	40±6 (26)	41±6 (4)	40±6 (22)	1.03
W. African[d]	36±12 (19)	35 (1)	38±11 (18)	0.92
Indian[e]	64±15 (6)	44 (1)	68±12 (5)	0.65
Australian[f]	40±8 (16)	38±8 (8)	42±8 (8)	0.90
Brazilian[g]	52±11 (12)	51±18 (2)	52±10 (10)	0.98

[a] Superior province, reference 6 with additional data not yet published; [b] Churchill province, 1; [c] 5,7,8,9,10,11; [d] 12; [e] 2; [f] 13, 14, 15; [g] 16. * – heat flow values adjusted for glaciation effects.

HEAT FLOW AND HEAT GENERATION IN GREENS. BELT

DRURY, M.

Although it appears from column 4 of Table 1 that mean heat flow in greenstone belts is indeed lower than that in crystalline areas of the shields, there is, in all shields except one, considerable overlap of the two values. The exception is the Indian Shield, but there is only one value from a greenstone belt for that. Further, in most cases no statistical signficance can be inferred as there are fewer data for greenstone belts than for crystalline areas. Taking the mean values, the heat flow from crystalline areas is apparently approximately 10% higher than that from greenstone belts.

Not all heat flow data used for compiling Table 1 had associated heat generation data. The most complete set is for the Canadian shield (Superior and Churchill provinces). Linear least squares regression for those data yields:

	Q_0 (mW/m^2)	D km	r	Q^* (mW/m^2)	A^* (μW/m^3)	n
Greenstone belts	33±4	7±6	0.45	37±7	0.51±0.46	7
Crystalline areas	26±6	12±4	0.67	40±9	1.16±0.51	11

where n is the number of data pairs, r is the correlation coefficient, Q^* is the mean heat flow and A^* is the mean heat generation of borehole samples. The correlations are low and statistically the differences between the parameters for the two crustal types are insignificant. However, assuming that radiogenic elements are distributed uniformly with depth to D, the value of D for the greenstones suggests that they are approximately 7 km thick, a value compatible with those cited by Condie (17). The data also suggest that the heat flow – heat generation relationship for the greenstones could be written as

$$Q = Q_0 = (D_c-D_g)A_c + D_gA_g \quad [2]$$

in which subscripts g and c refer to greenstone and crystalline crust and Q_0 is the reduced heat flow for the crystalline crust. This can be seen by inserting appropriate values for greenstones and crystalline terrain into equation [2]. It implies that the greenstones are underlain by normal crystalline crust, including 5 km of upper crust, but that they are not allochthonous, replacing 7 km of that crust rather than simply overlying it.

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THE WESTERN WABIGOON SUBPROVINCE, SUPERIOR PROVINCE, CANADA: LATE
ARCHEAN GREENSTONE SUCCESSION IN RIFTED BASEMENT COMPLEX.

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The Wabigoon Subprovince, interposed between the
predominantly metasedimentary-plutonic and gneissic English River
and Quetico Subprovinces to the north and south respectively,
exposes Archean greenstone and granitoid rocks for a strike
length of greater than 700 km. Based on predominating rock types,
the western part of the subprovince is divided into two terranes:
the northwestern Wabigoon volcano-sedimentary and plutonic
terrane (NWW) and the Wabigoon Diapiric Axis terrane (WDA)(1).
NWW in Ontario extends southwesterly from Savant Lake to
Lake of the Woods. Organized searches for older and younger age
limits for the evolution of this terrane, yield reliable zircon
U-Pb ages for supracrustal strata that span from 2755 Ma to 2711
Ma, although most ages are between 2720 Ma and 2734 Ma
(2,3,4,5,6,7). The lowermost volcanic sequence in the western
part of NWW is bimodal Mg-rich tholeiitic basalt and rhyodacite
at Thundercloud Lake (2755 Ma); later, at 2734 to 2718 Ma,
bimodal Fe-rich tholeiitic basalt and rhyodacite (Dash Lake) is
attended by bimodal basalt and tonalite plutonism. This stage
overlaps with intermediate to felsic calc-alkaline volcanism
(Kakagi Lake). The latest volcanism in the sequence at 2711 Ma is
dacite at Stephen Lake (3,7) which is conformable with the
subjacent Kakagi Lake strata and as such gives an upper limit for
the age of major tectonism affecting the supracrustal rocks.
WDA is a 400 km long by 75 km wide domal structure which
consists of 1) gneissic tonalitic to granodioritic rocks forming
domes and lesser massive segregations, 2) crescentic dioritic to
granitic plutons occurring at or near the contact between the
gneiss domes and the Wabigoon supracrustal rocks, and 3) later
plutons of diorite to granite (1,8,9). U-Pb geochronology
indicates that at least some of the eastern part of the terrane,
which extends from Steep Rock Lake in the south to Caribou Lake
in the north, has some old (approx. 3.0 Ga) gneissic and
supracrustal rocks (10). The western part of WDA, so far has not
yielded old ages; gneissic to massive tonalitic rocks have
intrusive ages of 2720-2725 Ma (3,7,9). At least some of the
gneissic tonalite forming the domes in the western part of WDA
have ages similar to, and in the field are gradational with,
tonalite plutons intruding NWW. A sphene U-Pb age of 2674 Ma for
gneissic tonalite with a zircon U-Pb age of 2723 Ma suggests that
the gneissification was a late event involving the resetting of
the sphene age but that the age of intrusion was retained by the
zircon. The crescentic and later plutons dated so far have ages
near 2700 Ma (3,7,9) and do not have regional foliation thus
providing an approximate lower limit for the age of major
tectonism in the terrane.
NWW is interpreted to have formed during rifting of a
basement complex that underlies the adjacent English River

Subprovince (11) and the western part of NWW and WDA. The complex
is approximately 3.0 Ga old and perhaps older. The rifting
started with mafic magmatism which evolved to be bimodal basalt-
rhyodacite. Tonalite intrusions accompanying the bimodal
volcanism caused little or no deformation of the adjacent
supracrustal rocks (12). Much of the contemporaneous
calc-alkaline sequence may be from mixing of basalt and tonalitic
magmas. The age of major deformation in the supracrustal rocks
may be bracketed by the age of the uppermost (and conformable)
Stephen Lake dacite at 2711 Ma and the age of the posttectonic
plutons at approximately 2700 Ma. Heating of the lower crust by
ponding of mafic magma caused most of the deformation of both the
younger Wabigoon 'rift' sequence and the basement complex; WDA is
the scar of maximum crustal diapirism transecting the new and old
crust.

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KOMATIITE GENESIS IN THE ARCHAEN MANTLE, WITH IMPLICATIONS FOR THE TECTONICS OF ARCHAEN GREENSTONE BELTS;   D. Elthon, Lunar and Planetary Institute, 3303 NASA Rd.1, Houston, TX 77058 and Department of Geosciences, University of Houston — University Park, Houston, TX 77004.

The presence of ultramafic lavas (komatiites) associated with Archaen greenstone belts has been suggested to indicate very high increments (50-80%) of partial melting of the Archean mantle [e.g., 1-3]. Such extensive melting of the Earth's mantle during the Archaen might have profound effects on the early tectonic and chemical evolution of the planet [e.g., 4 & 5], although problems associated with keeping the komatiite liquid in equilibrium with the residual mantle at such high increments of melting has cast doubt upon aspects of extensive melting [e.g.,6 & 7]. Two important aspects of the origin of komatiites are discussed below.

I. WHAT IS THE NATURE OF PRIMARY KOMATIITE LIQUIDS?
    One of the most fundamental aspects of understanding the tectonic and geochemical mode of origin for komatiites is the problem of komatiite primary magmas. The identification of primary komatiite magmas is complicated by the extensive metamorphism that these rocks have typically undergone and by olivine (+ minor spinel) crystallization at low pressures (~1 atm). The crystallization of olivine rapidly depletes a komatiite liquid in MgO, such that the most likely candidates for primary magmas are those with the highest MgO contents.

    Previous efforts to evaluate primary komatiitic liquids have proposed that they might contain as much as 33% MgO [2] or 30% MgO [8]. These studies have relied principally on comparison of the compositions of olivines crystallized in high-pressure experimental studies of komatiites with relict olivines found in komatiites (as high as ~Fo94).

    The Fe-Mg exchange between olivine and basaltic-komatiitic liquids has recently been summarized by [9], in which they present equations for calculation of olivine-liquid equilibria over a wide temperature (1074-1600° C.) and pressure (1 atm to 25 kbar) range. The KD values for a wide range of komatiites (>20% MgO) were calculated using this equation and range from 0.28 to 0.31 at temperatures of 1450-1650°C at 1 atm. This olivine-liquid equilibrium is shown in Fig. 1 along with the compositions of the most magnesian olivines in komatiites (olivine and komatiite compositions from [8] and references therein). The 1 atm KD values have been used here because the present author considers it most likely that the olivines in komatiites have crystallized at very low pressures (~1 atm); previous investigators [2 & 4] have used KD values from high pressure experiments, which are substantially higher [10 & 11].

    The data shown in Fig. 1 (horizontal lines connect the olivine compositions with the liquid from which they could have crystallized) indicate that the komatiite olivines probably have crystallized from liquids with Fe/Mg >0.230. This Fe/Mg (0.230) corresponds to 22-25% MgO in the komatiites, depending upon the FeO content of the liquid. This data indicates that the most magnesian olivines in komatiites could have crystallized from liquids with 22-25% MgO, in contrast to previous estimates of 30-33% MgO. These liquids will have liquidus temperatures of ~1500°C at 1 atm pressure. More MgO-rich komatiites have probably become enriched in MgO as a consequence of olivine accumulation and/or Mg metasomatism.

Fig. 1
Cation Fe/Mg
in olivine
versus cation
Fe/Mg in the
liquid. The
horizontal
lines connect
the composition
of olivines
in komatiites
with liquids
in equilibrium
with these
olivines.

Cation Fe/Mg in Liq.
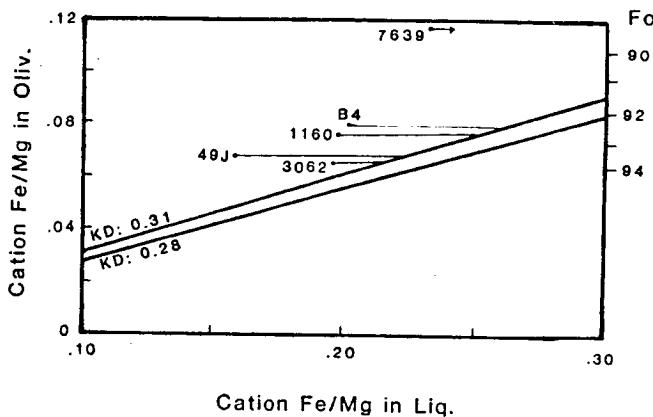
II. WHAT PERCENTAGE OF MELTING IS REQUIRED TO PRODUCE KOMATIITES?

As noted above, it is generally assumed that a very high increment
of melting (50-80%) is required in order to generate komatiites from the
Earth's mantle. Experimental studies of the melting of reasonable mantle
compositions have shown that very magnesium-rich magmas may be produced at
high increments of melting [e.g., 11-13]. As outlined below, however,
these MgO-rich magmas produced by very large increments (40-80%) of melting
within the mantle are NOT komatiites.

A pseudo-liquidus phase diagram for evaluating the petrogenesis of
komatiites is shown in Fig. 2. At low pressures (1 atm), some magmas
(those above the Ol-R join) will crystallize augite as the first pyroxene
and others (those below the OL-R join) will crystallize pigeonite or
orthopyroxene first. Field and petrographic studies of komatiites have
shown that they crystallize augite as the first pyroxene in virtually all
instances. Most terrestrial magmas also crystallize augite as the first
pyroxene; boninites are an obvious exception. Also shown in Fig. 2 is the
field for the compositions of komatiites from Munro Township, which
crystallize augite as the first pyroxene [14].

A partially schematic melting path for melting of the mantle is shown
in Fig. 2 for melting at 15 kbars. At small to moderate increments of
melting (<30%), the primary liquids will lie above the OL-R join, but will
lie below the OL-R join at larger increments of melting. The extent of
melting required to produce primary magmas below the OL-R join will vary as
a function of the composition of the mantle and the pressure of melting,
but it is clear that high increments of melting that might produce dunite
or OPX-poor harzburgite residues will produce primary magmas that will lie
below the OL-R join and will evolve to crystallize orthopyroxene and/or
pigeonite before augite. The extent of melting most likely to produce
komatiitic magmas is more like 20-25% rather than the 50-80% previously
proposed. Although not discussed by the previous authors, this feature is
further apparent in the data of [12 & 13].

Spinels from komatiites have Cr/(Cr+Al) from ~0.70 to 0.80 [15],
which would suggest a slightly higher increment of partial melting of the
mantle that occurs in the present-day suboceanic mantle [16], rather than
the much higher increments proposed in previous studies.

Fig. 2
Pseudo-liquidus phase diagram
projected from plagioclase[17].
Pseudo-invariant points at
1 atm (R), 10, 15, 20
and 25 kbars are
shown.
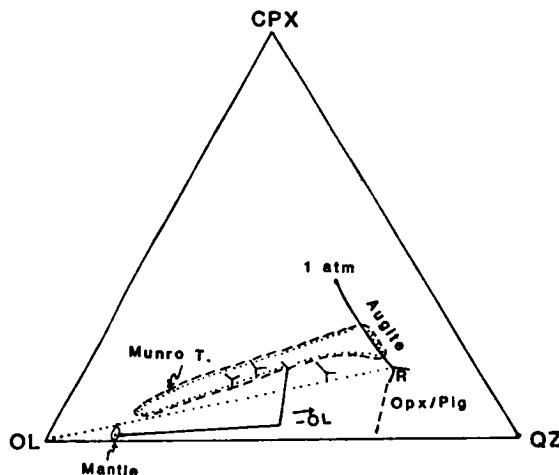
In summary, it is suggested that the extent of partial melting that produces komatiite primary magmas is ~20-25% and that these magmas have 22-25% MgO or less. This substantially lower estimate for the extent of melting and eruption temperatures will certainly influence those tectonic characteristics of greenstone belts associated with the dynamics of mantle upwelling and convection.

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THE YILGARN CRATON WESTERN AUSTRALIA : A TECTONIC
SYNTHESIS : I   R.E.P. Fripp, Western Australian Institute of Technology

The Yilgarn Craton in Western Australia is one of the larger contiguous preserved Archaean crustal fragments, with an area of about 650,000 square kilometres. Of this, by area, about 70% is granitoid and 30% greenstone. The Craton is defined by the Darling Fault on its western margin, by Proterozoic deformation belts on its southern and northwestern margins, and by unconformable younger sediments on its eastern and northeastern margins.

A regional geotectonic synthesis at a scale of 1:500,000 is being prepared. This is based largely upon the 1:250,000 scale mapping of the Geological Survey of Western Australia together with interpretation using geophysical data, mainly airborne magnetic surveys.

On a regional basis the granitoids are classified as pre-, syn- and post-tectonic (1) with respect to greenstone belt deformation. The post-tectonic granitoids yield Rb-Sr isochrons of about 2.6 b.y., close to Rb-Sr ages for the greenstones themselves which are up to about 2.8. b.y. old (2), although data for the latter is sparse.

Contacts between earlier granitoids and greenstones which are not obscured by the post-tectonic granitoids are most commonly tectonic contacts, intensely deformed and with mylonitic fabrics. The general concensus however is that there is a pre-tectonic, pre-greenstone sialic gneiss preserved in places (1,3).

Existing models for the evolution of the belts involve 3 large basinal structures ("broad elongate downwarps"), of which the Eastern one (the Noreseman-Wiluna Belt) is considered to be a rift fault-bounded graben (1). The postulated basins are separated by large tabular belts of discordant post-tectonic granite when viewed regionally. This may be a 'red herring'. It is possible that, for example, the entire greenstone package preserved on the Craton was part of one basin, or numerous combinations and parts of basins. There is no compelling diagnostic evidence collated to date to postulate on the original disposition, geometry and relationships between belts.

This synthesis is a preliminary attempt at addressing this problem, by attempting to decipher the broad tectonic-stratigraphic sequences preserved and thereby to reconstruct, as far as is possible, the original nature of the greenstones. There is structural evidence to suggest that the deformation histories of the greenstones and some of their surrounding and occluded granitoids involves early fold-nappe tectonics in places, and possibly thrust nappes, as well as late large-scale imbrication or slicing. During early deformation of the belts, massif-style nappe tectonics may have occurred in places, on scales not dissimilar to those seen in young fold belts.

It is intended, with future work, to test these postulates and to examine whether the tectonic history of the Yilgarn Craton is indicative of the loss of considerable greenstone (back to the womb?) and perversely (sic), its local preservation by obduction and stacking. How well can we reconstruct the deformed granitoids and greenstones, in their undamaged state?

THE YILGARN CRATON WESTERN AUSTRALIA
Fripp, R.E.P.

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TECTONICS OF SOME AMAZONIAN GREENSTONE BELTS; Allan K. Gibbs, Institute for the Study of the Continents, and Dept. Geol. Sci., Cornell University, Ithaca, New York, 14853

Greenstone belts exposed amid gneisses, granitoid rocks, and less abundant granulites along the northern and eastern margins of the Amazonian Craton yield Trans-Amazonican metamorphic ages of 2.0-2.1 Ga. (Regional geology: 1-13). Early Proterozoic belts in the northern region probably originated as ensimatic island arc complexes. The Archean Carajas belt in the southeeeastern craton probably formed in an extensional basin on older continental basement. That basement contains older Archean belts with pillow basalts and komatiites. Belts of ultramafic rocks warrant investigation as possible ophiolites.

NORTHERN BELTS - Volcanic rocks of the northern belts were erupted in the Early Proterozoic (2.3-2.1 Ga)(14-17). The contiguous belts of Guyana (18,19) and Venezuela (20,21) closely resemble those of Suriname (7-9,22) and French Guiana (1,4,16,23), though the two regions are separated by the Central Guiana Granulite Belt. Typical sections consist of a lower flow and pillowed low-K basalt-gabbro unit, overlain by interbedded mafic, intermediate, and felsic volcanics of both tholeiitic and calc-alkaline suites; overlain by and interstratified with volcaniclastic greywackes, pelites, and chemical sedimentary rocks. Basalts with pronounced iron-enrichment and others with high magnesium contents are both present, as are both tholeiitic and calc-alkaline andesites and felsic volcanics (18,19,22,24). Generally conformable tuffaceous and epiclastic conglomerates, greywackes, lithic arenites, and shales appear petrographically and geochemically to have been derived from the associated volcanic rocks, without significant contributions from continental sources (18,25). The relative abundances and types of volcanic and sedimentary rocks vary: felsic volcanics are irregularly distributed, and magnesian basalts and possible komatiites are particularly common in central French Guiana (22). Ultramafic, mafic, and anorthositic intrusive complexes may be genetically associated with some of the volcanic rocks (1,18,23). Some belts are overlain by quartz-rich epiclastic sedimentary rocks that were folded and metamorphosed with the belts but appear to be unconformable (1,13).

The northern belts have randomly-branching synclinal map patterns. Prominent metamorphic foliations generally correlate with the regional folds, with foliations locally crenulated or destroyed by younger shear deformation, which elongated (WNW-ESE) both the belts and associated granitoid rocks. Metamorphic grades range from amphibolite on the belts' peripheries to lower greenschist and zeolite in the interiors. Diverse local mineral assemblages indicate high, intermediate, and low-pressure metamorphic series. Anatectic, two-mica granites intrude metapelitic schists along the northern periphery.

No evidence has been reported of basement-cover relations between the northern belts and adjacent gneisses. Field observations and geochemical similarities suggest that the greenstones pass into the intervening gneisses by increase in metamorphic grade (14,15,17,26-28). The associated granulites also appear to represent Early Proterozoic, rather than Archean crust (16,27,29). Sm-Nd and Rb-Sr isotopic systematics indicate that little if any older continental crust was involved in this greenstone-belt volcanism.

The northern belts are thought to have been originally contiguous with the Birimmian belts of west Africa. Mature sedimentary rocks overlying the greenstone belts have much in common with the Tarkwaian of West Africa.

EASTERN BELTS - Belts of the east-central craton (30,31) have not been adequately dated. Most lithostratigraphic sections have not yet been resolved, in part due to intense deformation and common medium grade metamorphism. Prominent banded iron formations, ultramafic schists, and current-bedded, fuchsite-bearing quartz arenites and conglomerates are present: these lithologies are uncommon in the northern belts. Small enclaves of iron formations and chromite-bearing ultramafic rocks occur in south and central Suriname, and might correlate with the east-central belts.

Archean greenstone belts with pillow basalts and komatiites, and belts of serpentinite occur amid granitoid rocks and gneisses in the southeastern craton, apparently forming a basement to the Serra dos Carajas belt (32). The latter has a dominantly mafic bimodal volcanic suite, roughly 4-6 km thick and dated at 2.75 Ga, overlain by 100-300 m of iron formation, and a 1-2 km thick fine clastic and chemical sedimentary complex (33,34). The mafic rocks are unlike typical Archean basalts and basaltic andesites, but have chemical and isotopic evidence of contamination with older continental crust, like many basalts of modern contiental extensional settings.

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AMAZONIAN GREENSTONE BELTS
Gibbs, A. K.
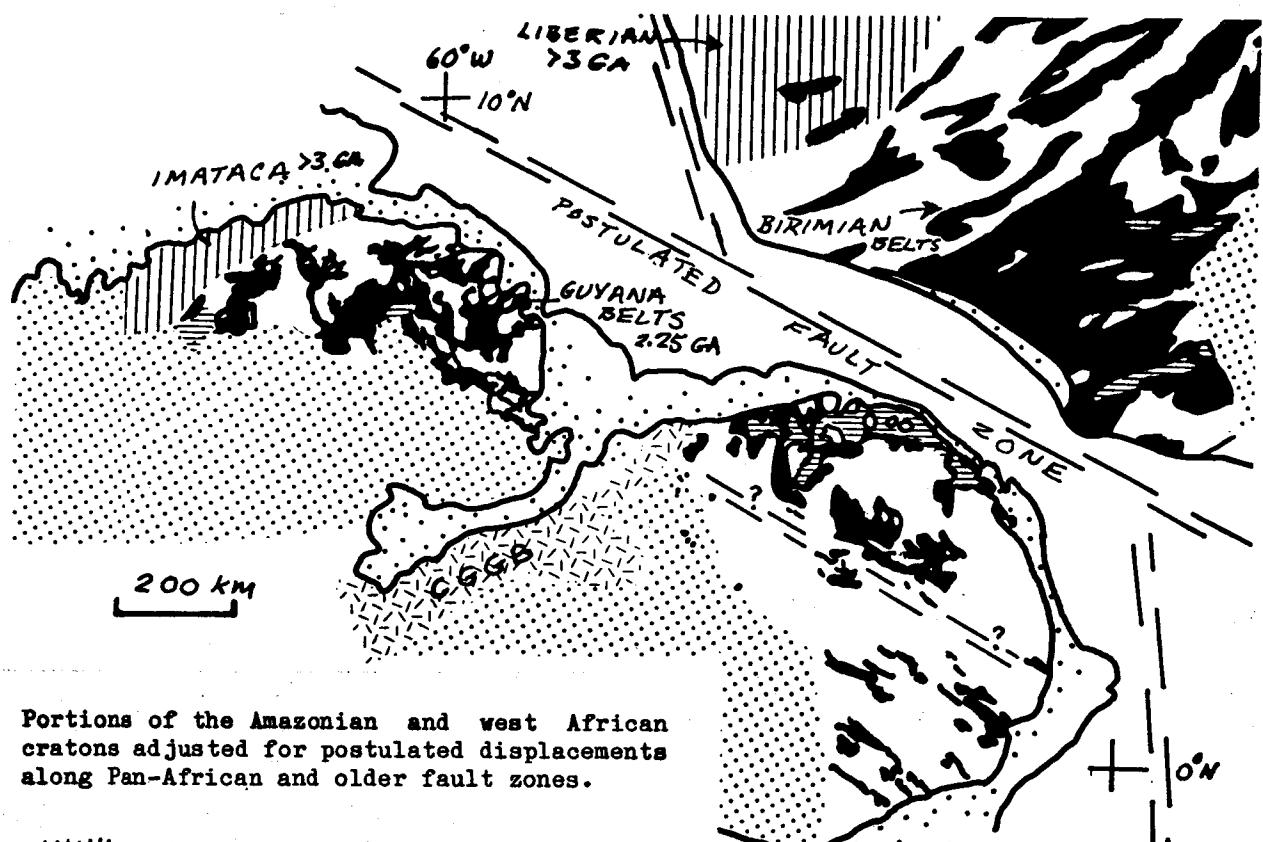
Portions of the Amazonian and west African cratons adjusted for postulated displacements along Pan-African and older fault zones.

||||||| - Archean Imataca and Liberian terranes

- Central Guiana Granulite Belt <2.5 Ga

- greenstone belts (sensu lato)

≡ - quartz-rich metasedimentary rocks considered unconformable on the greenstone belts

- granitoid rocks and gneisses

:::::: - areas with abundant continental igneous and sedimentary cover <1.9 Ga

Postions of the cratons (13, based in part on 35) is compatable with paleomagnetic data (36), and juxtaposes geolgical features in the two cratons. A major geological province boundary must be present in the Amazonian craton between the Carajas belt and the northern belts (17,37,12): one possible position is shown.
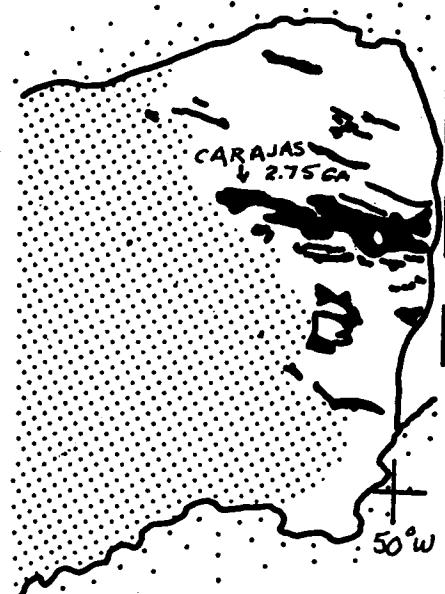
Spatial greenstone-gneiss relationships:
evidence from mafic-ultramafic xenolith
distribution patterns

A.Y. Glikson

Division of Petrology and Geochemistry, Australian
Bureau of Mineral Resources, Geology and Geophysics

ABSTRACT

The distribution patterns of mafic-ultramafic xenoliths
within Archaean orthogneiss terrain furnish an essential
key for the elucidation of granite-greenstone relations.
A complete gradation in scale exists between synclines,
large-scale outliers and outcrop-scale xenoliths of mafic
and ultramafic metavolcanic rocks. Accordingly, most
greenstone belts constitute "mega-xenoliths" rather than
primary basin structures. Transition along strike and
across strike between stratigraphically low greenstone
sequences and xenolith chains demonstrate their contempo-
raneity, as shown for example in Fig. 1 where the relati-
onships between the Holenarsipur greenstone belts and
associated xenoliths in southern India are portrayed.
Regional to mesoscopic-scale characteristics of xenolith
swarms and their relations with early greenstone units
are well expressed in parts of the Pilbara Block, Western
Australia. Xenolith distribution patterns in dome-
arcuate syncline gneiss-greenstone terrains define subsi-
diary gneiss domes within the batholiths. These terrains
represent least deformed cratonic "islands" within an
otherwise penetratively foliated deformed gneiss-green-
stone crust. The oval gneiss domes are thought to have
developed originally by magmatic diapirism - evidenced by
intrusive relations and contact aureoles - followed by
late-stage solid state uprise related to isostatic adjust-
ments. The late vertical movements were associated with
development of major shear zones along tectonized boun-
dary zones of batholiths, where interdigitated deformed
gneiss-amphibolite schist intercalations were derived by
the attenuation of xenolith-rich orthogneiss. The defor-
mation process involved interthrusting and refolding of
the interleaved plutonic and supracrustal units. The
exposure of high grade metamorphic sectors is related to
uplift of deep seated zones of the batholiths along
reactivated faulted boundaries. Transitions from granite-
greenstone terrains into gneiss-granulite suites involve
a decrease in the abundance of supracrustal enclaves and
an increased strain rate. Whereas early greenstone
sequences are invariably intruded by tonalitic/trondhje-
mitic/granodioritic gneisses, stratigraphically higher
successions may locally overlap older gneiss terrains and
their entrained xenoliths unconformably. The contiguity

SPATIAL GREENSTONE-GNEISS RELATIONSHIPS
A. Y. Glikson

of xenolith patterns suggests their derivation as relics
of regional mafic-ultramafic volcanic crustal units and
places limits on horizontal movements between individual
crustal blocks.
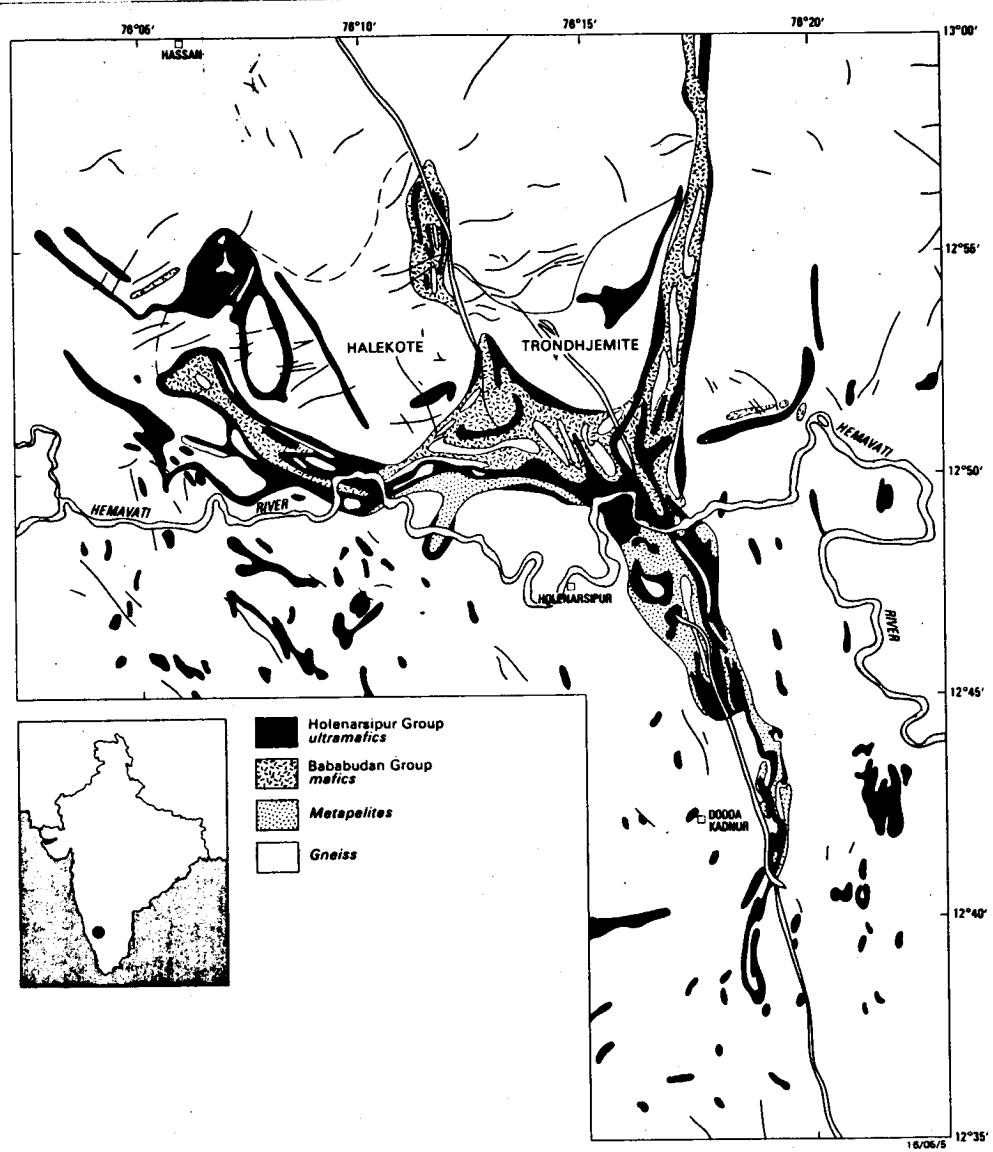
Fig. 1 - A geological sketch map of the Holenarsipur
greenstone belt, Karnataka (after Naqvi, 1981, J. Geol.
Soc. India, 22:458-469)

SPATIAL GREENSTONE-GNEISS RELATIONSHIPS
A. Y. Glikson

Alternative models of granite-greenstone relations are
portrayed in Fig. 2. Model 1 applies to late greenstone
belts overlapping sial whereas model 2 to early belts or
stratigraphically basal volcanic units believed to be
derived from simatic crust. Major detachments along
gneiss-greenstone boundaries and local overfolding and
thrusting suggest horizontal tectonic translations.
These are overprinted by the dominantly vertical tectonic
movements related to the diapriric (magmatic and post-
magmatic) uprise of the tonalite/trondhjemite plutons.
The contiguous temporal-spatial grid outlined by the
xenolith swarms constrains major lateral movements of
individual blocks relative to each other, placing limits
on plate tectonics interpretations.
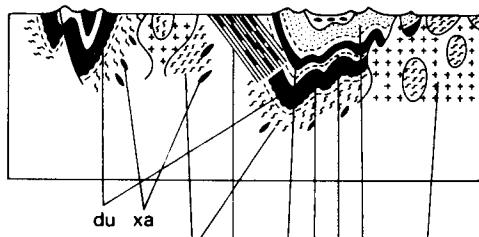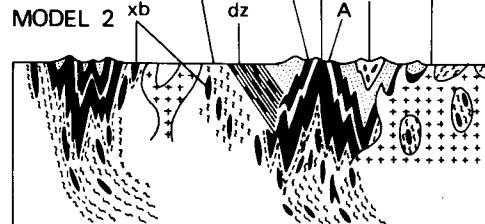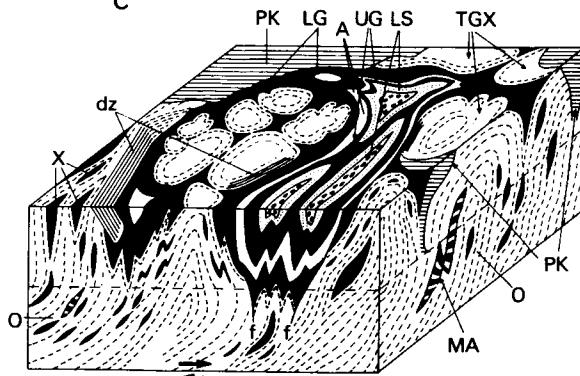
Fig. 2 - Alternative
models of gneiss-
greenstone relation-
ships.
a - model 1 - gneiss-
greenstone basement-
cover relations, invo-
lving deformed uncon-
formities (du).
b - model 2 - gneiss-
greenstone relations
involving primary and
deformed intrusive
contacts.
c - model 2 portrayed
in block diagram,
showing transition
from granite-greens-
tone to gneiss-granu-
lite terrain with
crustal depth.
LG - lower greenstone
UG - upper greenstone
A - acid volcanics &
sediments; TGX - Na-
gneiss with xenoliths;
PK - late granites;
dz - deformed zone;
LS - late sediments;
O - orthogneiss;
MA - mafic and anortho-
sitic inclusions;
x, xa, xb - xenoliths

DISMEMBERED ARCHEAN OPHIOLITE IN THE SOUTHEASTERN WIND RIVER
MOUNTAINS, WYOMING -- REMAINS OF ARCHEAN OCEANIC CRUST. G.D. Harper,
Dept. Geological Sciences, State University of New York, Albany, NY 12222

Archean mafic and ultramafic rocks occur in the southeastern Wind
River Mountains near Atlantic City, Wyoming (Figure) and are interpreted
to represent a dismembered ophiolite suite. The ophiolitic rocks occur
in a thin belt intruded by the 2.6 Ga Louis Lake Batholith on the
northwest (1, 2). On the southeast they are in fault contact with the
Miners Delight Formation comprised primarily of metagraywackes with minor
calc-alkaline volcanics.

The ophiolitic and associated metasedimentary rocks (Goldman Meadows
Formation) have been multiply deformed and metamorphosed. The most
prominant structures are a pronounced steeply plunging stretching
lineation and steeply dipping foliation. Pillow lavas are stretched
parallel to the lineation and typically have aspect ratios of 10:3:1.
Bedding in banded iron formation shows polyphase folding with fold axes
parallel to the stretching lineation; sheath folds are locally well
developed. The intrusive contact of the Louis Lake batholith with the
ophiolitic rocks has been extensively modified by deformation; the
batholith becomes progessively more deformed as the contact is
approached, and at the contact the batholith is strongly lineated and
mylonitic. The contact between the ophiolitic rocks and the Miners
Delight Formation is a major fault zone (Roundtop Fault) containing
amphibolite-facies mylonites overprinted by greenschist-facies brittle
cataclasites (3). These structural data indicate that the ophiolitic and
associated metasedimentary rocks have been deformed by simple shear when
the Miners Delight was emplaced over the Louis Lake batholith and its
ophiolitic wall rocks.

The ophiolitic rocks include ultramafics, metagabbros, metadiabases,
and pillow lavas. Relict structures and textures are often well
preserved. However, an ophiolite "stratigraphy" is not present; the
ophiolitic rocks consist of tectonic slices, from northwest to southeast,
of (1) metadiabase, (2) metagabbro and ultramafics, (3) pelitic schists,
quartzite, and banded iron formation (Goldman Meadows Formation), and (4)
greenschist and amphibolite (Roundtop Mountain Greenstone) locally
containing pillows and massive flows or sills. In addition, a thin
sliver of pillow lavas occurs between the metadiabase and ultramafic
rocks at one locality, but is separated from the metadiabase by a
strongly foliated talc-actinolite-chlorite schist.

The ultramafic rocks are largely serpentinites, but some have
amphibole-chlorite assemblages and one clinopyroxenite was found. Many
of the ultramafic rocks and associated metagabbros have well-preserved
relict cumulus textures, and igneous layering is visible in a few
outcrops. The ultramafic rocks and associated metagabbros are only
weakly deformed, in contrast to the highly deformed mafic and
metasedimentary rocks.

Metadiabase occurs in a wide belt along the margin of the Louis Lake
batholith, and much of it occurs as large xenoliths within the margin of
the batholith. The metadiabase unit locally contains numerous parallel
dikes, some of which show one-way chilling. Medium to coarse-grained

ARCHEAN OPHIOLITE
G.D. Harper

metagabbro occurs locally within the metadiabase; some of the metagabbro occurs as thin screens between fine-grained metadiabase dikes. These features suggest that the metadiabase unit represents a deformed sheeted dike complex.

The Roundtop Mountain Greenstone contains common pillow structures with well-preserved chilled rims. Massive lavas or sills comprise a significant portion of the formation, and gray phyllites occur rarely.

Rare isolated outcrops of black and foliated "basaltic komatiites," consisting primarily of actinolite and chlorite, occur in both the Roundtop Mountain Greenstone and metadiabase. However, they are chemically very different from the pillow lavas and metadiabases and possibly represent younger alkaline dikes.

Metasedimentary rocks of the Goldman Meadows Formation overlying(?) the Roundtop Mountain Greenstone consist of pelitic schist, quartzite, and banded iron formation (1). The banded iron formation possibly formed by precipitation from hydrothermal vents in a manner similar to modern metalliferous sediments formed at spreading centers (4). Mafic sills and dikes (amphibolites) intrude the metasedimentary rocks, and are themselves deformed and metamorphosed.

Geochemical analyses were made of the metadiabase and pillow lavas to determine whether they are genetically related (5). "Immobile" trace element compositions (Ti, V, Cr, Ni, Zr, Y, Nb) are very similar in both units, consistent with the interpretation that they comprise different parts of a dismembered ophiolite. These rocks are similar to modern enriched mid-ocean ridge basalts.

The ophiolitic rocks are interpreted as the remains of Archean oceanic crust, probably formed at either a mid-ocean ridge or back-arc basin. All the units of a complete ophiolite are present except for upper mantle peridotities. The absence of upper mantle rocks may be the result of detatchment within the crust, rather than within the upper mantle, during emplacement. This could have been the result of a steeper geothermal gradient in the Archean oceanic lithosphere, or may have resulted from a thicker oceanic crust in the Archean (6).

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ARCHEAN OPHIOLITE
G.D. Harper

LOUIS LAKE BATHOLITH
Granodiorite

GOLDMAN MEADOWS FORMATION
Metasedimentary Rocks
(BIF = Banded Iron Formation)
BIF

ROUNDTOP MTN. GREENSTONE
Pillow Lava

METADIABASE
md-gr = metadiabase extensively
intruded by granodiorite
md | md-gr

METAGABBRO

ULTRAMAFIC ROCKS

MINER'S DELIGHT FORMATION
Turbidites (shaded)
Meta-andesite (triangles)

Foliation and Lineation

WYOMING
MAP AREA
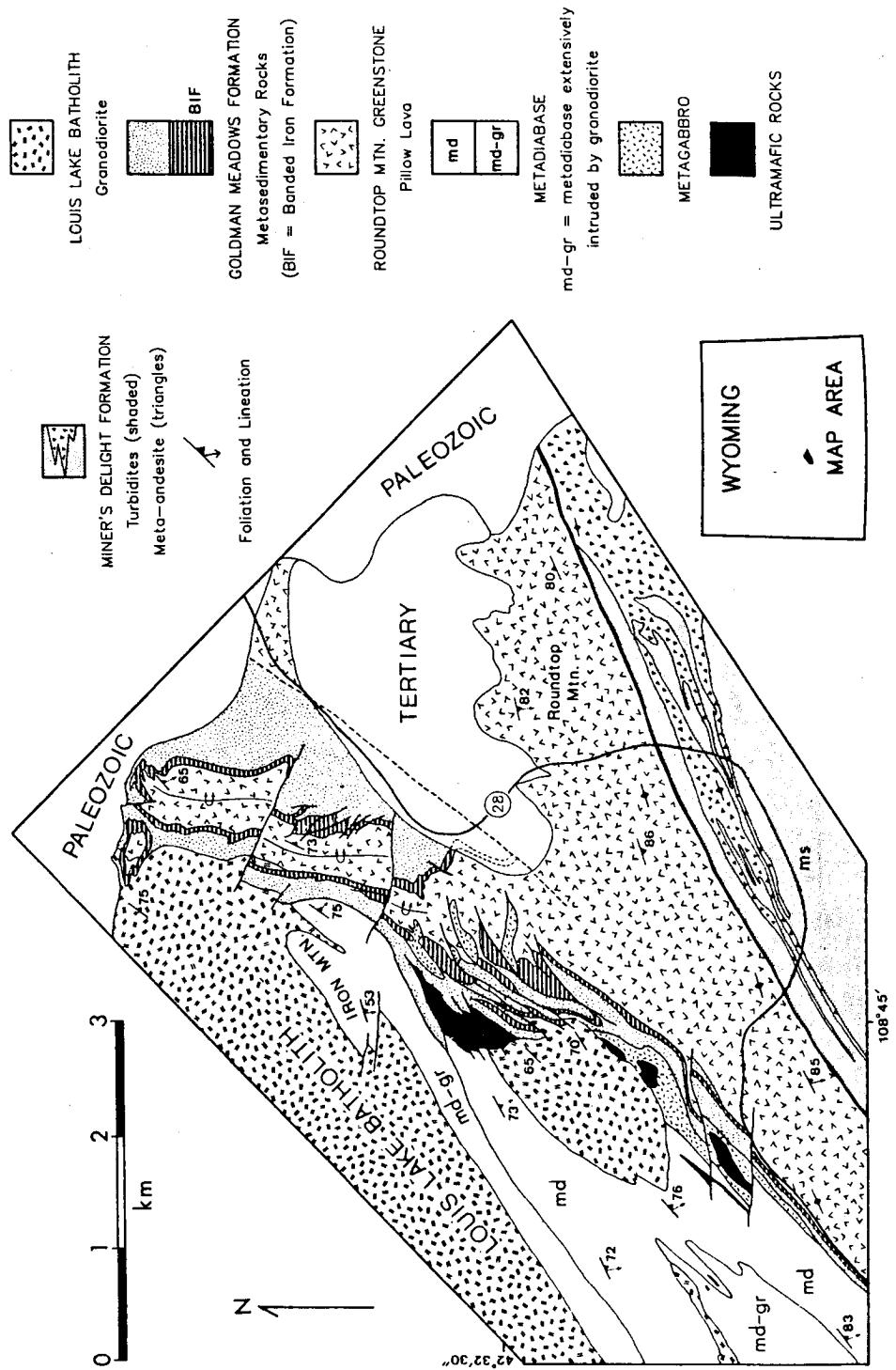
THE KOLAR SCHIST BELT: A POSSIBLE ARCHEAN SUTURE ZONE

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The Kolar Schist Belt in the Karnataka craton, south India, is a 4 to 20 km by 80 km long, N-S trending Archean supracrustal belt dominated by mafic metavolcanics. The schist belt is surrounded on both sides by granodioritic gneisses collectively known as the Peninsular Gneiss. Our work has shown that the Kolar Schist Belt and the surrounding gneisses include major discontinuities in age, structural style, and composition. These discontinuities are defined by the schist belt itself.

The results reported here are based on our Rb-Sr, Sm-Nd, and Pb-Pb whole rock isotope data; U-Pb dating of zircon and sphene; major and trace element (including REE) analyses; and field observations.

The schist belt is broadly synformal, but is complexly refolded into basin and dome structures (D. Mukhopadyay, personal communication). The first period involved N-S trending isoclinal recumbent folds during E-W compression. These folds were refolded into tight, upright folds along E-W trending axes. This sequence is broadly similar to those seen in other schist belts in the western part of the Karnataka craton.

Contacts between the Peninsular Gneiss and the margins of the belt have long been thought to represent an erosional unconformity. However, our recent field work indicates that the rocks at the contacts are physically interleaved by left lateral shearing. Due to this shearing the adjoining gneisses have been converted to quartz-muscovite schists, which were previously interpreted to be metasedimentary rocks.

The gneisses east of the schist belt are relatively homogeneous, granodioritic gneisses which were folded prior to intrusion of minor felsic bodies. Folds have not yet been defined in these gneisses, but a strong foliation was developed which strikes NNE and dips steeply to the west, suggesting horizontal compression.

The gneisses west of the schist belt show a much more complex, earlier history than that of the eastern gneisses. The granodioritic Dod Gneiss is the earliest unit on the western side of the schist belt. This rock was subjected to a period of deformation shown by an early foliation seen in some less-strained exposures. Subsequently, the Dod Gneiss was intruded by the leucocratic, granodioritic Dosa Gneiss and the granodioritic Patna Granite.

Following the intrusion of the Dosa Gneiss, the terrane to the west of the schist belt was subjected to a period of horizontal compression producing tight to isoclinal, W overturned folds with gently N or S plunging axes. The strong NNE axial planar foliation produced by this deformation is cut by the later N-S shears along the western margin of the schist belt.

THE KOLAR SCHIST BELT
Hanson, G. N., et al.

The gneisses on the east and west side of the belt have been dated using U-Pb ages for small populations of abraded zircons and abraded single zircons as well as sphene. These zircons commonly give concordant ages, in which case the small populations of zircons (ca. 100 micrograms) have analytical uncertainties of less than 1 Ma. and the single zircons have uncertainties of about 5 Ma.

Gneisses east of the belt were intruded at 2529±1 Ma based on U-Pb ages for zircon. This age is consistent with the Rb-Sr and Pb/Pb whole rock isochron ages. The isochrons have a mantle-like initial ratio for Sr (87/86=0.7013) and mu=8 for the Pb data. These values suggest that the gneisses were not derived from a much older continental crust. U-Pb ages for metamorphic sphene are 2520 ±1 Ma suggesting that the gneisses were metamorphosed to at least amphibolite grade at that time.

West of the belt, based on U-Pb ages for zircon, the Dod Gneiss was emplaced at 2610±5 Ma, the Dosa Gneiss was intruded at 2550±10 Ma and the Patna Granite at 2551±1 Ma. The time of metamorphism based on the U-Pb ages for sphene from the Dod Gneiss is 2551±1 Ma. Rb-Sr and Pb/Pb whole rock data suggest that the gneisses were variably contaminated by an older basement. U-Pb ages for some of the single zircon cores from the Dod Gneiss and later aplitic dikes indicate a zircon component was inherited from this basement, which has a minimum age of 3200 Ma. The basement, which has not yet been clearly identified in the field, seems to include quite evolved felsic rocks.

In the Kolar Schist Belt there are two suites of komatiitic and tholeiitic amphibolites. Both the komatiitic and tholeiitic amphibolites on the eastern side are light REE enriched, and almost all of the komatiitic and tholeiitic amphibolites in the west-central part of the belt are lightest REE depleted. The preservation of rare pillow structures and the association of the amphibolites with iron formation suggest that the amphibolites were formed under submarine conditions. The grade of metamorphism is amphibolite facies.

Rajamani et al. (1) concluded that the komatiitic amphibolites from both the east and west central part of the belt were derived by 10 to 25% melting at depths greater than 80 km and at temperatures greater than 1500°C in a mantle with an FeO/MgO ratio greater than that of pyrolite. Other models proposed for the generation of komatiites generally require larger percentages of melting to generate the high MgO abundances.

Rajamani et al. (1 and in preparation) suggest that the tholeiites appear to have been derived by melting at shallower levels than the komatiites and derived from sources which were highly variable in their FeO/MgO ratios, generally with FeO/MgO ratios much greater than that for the sources for the komatiites. The key arguments are that: the tholeiites are very iron-enriched

compared to the field for potential melts of pyrolite at pressures less than 25 kb on an olivine saturation surface; and while the incompatible elements show similar ratios in the komatiites and tholeiites for each suite, the expected correlations between major and trace elements for differentiation from komatiites or melting of sources similar to those of komatiites are not found.

Sm-Nd data for komatiites from both sides of the belt lie with large variations about a 2900 Ma isochron. It is not clear why the data lie about a 2900 Ma isochron. Is this the age of these amphibolites? If this is so, they are much older than the igneous felsic rocks on either side of the belt which are 2500 to 2600 Ma. Or, is this the time when the sources became variably light REE enriched and depleted? Some of the variation in the Sm/Nd ratios is clearly a function of melting processes in which garnet was left in the residue. Perhaps the variability in the data about the reference line reflects a number of reasons such as: variable times of light REE depletion and enrichment of their mantle sources; as well as the possible effects of crustal contamination or metamorphic alteration.

Even though the ages of the units making up the Kolar Schist Belt are poorly constrained, the sources of the amphibolites so far analyzed had long-term histories of LREE depletion (epsilon Nd of +2 to +8 for an age of 2900 m.y.) relative to other Archean mafic rocks which commonly have epsilon Nd equal to about +2.0 ± 2.0.

The Kolar Schist Belt represents a N-S trending discontinuity in the structures, lithologies, and emplacement and metamorphic ages of late Archean gneisses. The suggestion of a much older basement on the west side of the belt is not seen on the east. Within the schist belt amphibolites from each side have distinctly different chemical characteristics, suggesting different sources at similar mantle depths. These amphibolites were probably not part of a single volcanic sequence, but may have formed about the same time in two completely different settings. Could the amphibolites with depleted light REE patterns represent Archean ocean floor volcanics which are derived from a mantle source with a long term depletion of the light REE? Why are the amphibolites giving an age which may be older than the exposed gneisses immediately on either side of the belt? These results suggest that it is necessary to seriously consider whether the Kolar Schist Belt may be a suture between two late Archean continental terranes.

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PRELIMINARY REPORT ON THE GEOLOGY AND GOLD MINERALIZATION OF THE SOUTH PASS GRANITE-GREENSTONE TERRAIN, WIND RIVER MOUNTAINS, WESTERN WYOMING (USA); W.D. Hausel, Geological Survey of Wyoming, Laramie, Wyoming 82071

The South Pass granite-greenstone terrain lies near the southern tip of the Wind River Mountains of western Wyoming. This Archean supracrustal pile has been Wyoming's most prolific source of gold and iron ore. From 1962 to 1983, more than 90 million tons of iron ore were recovered from oxide-facies banded iron formation, and an estimated 325,000 ounces of gold were mined from metagreywacke-hosted shears and associated placers (1).

Precambrian rocks at South Pass are unconformably overlain by Paleozoic sediments along the northeast flank, and a Tertiary pediment buries Archean supracrustals on the west and south. To the northwest, the supracrustals terminate against granodiorite of the Louis Lake batholith; to the east, the supracrustals terminate against granite of the Granite Mountains batholith. The Louis Lake granodiorite is approximately 2,630 $\pm$ 20 m.y. old (2), and the Granite Mountains granite averages 2,600 m.y. old (3).

The geometry of the greenstone belt is best expressed as a synform that has been modified by complex faulting and folding. Metamorphism is amphibolite grade surrounding a small island of greenschist facies rocks.

The youngest of the Archean supracrustal successions is the Miners Delight Formation. This unit yielded a Rb-Sr isochron of 2,800 m.y. (2). A sample of galena from the Snowbird Mine within the Mienrs Delight Formation yielded a model age averaging 2,750 m.y. (4). The Snowbird mineralization appears to be syngenetic and is hosted by metavolcanics of calc-alkaline affinity.

Based on regional mapping by Bayley and others (5) and by the author (in progress), four mappable supracrustal units are present. The uppermost unit, the Miners Delight Formation is greater than 1,600 m thick and consists of metagreywacke, metavolcanics, metaconglomerate, graphitic schist, and tremolite-actinolite schist. Underlying, and in fault contact with turbidites in the Miners Delight Formation, are metatholeiites of the Roundtop Mountain Formation. These metatholeiites are amphibolites, greenstones, and pillow metabasalts. The geometry of the pillows, which has been used for determining the tops and bottoms of units (5, 6) has only produced ambiguous conclusions due to the intense deformation.

The Roundtop Mountain greenstones are underlain(?) by quartzite, metapelite, and banded iron formation of the Goldman Meadows Formation. This unit, in turn, is underlain(?) by mafic and ultramafic schists tentatively named the Diamond Springs ultramafics. This ultramafic unit consists of amphibolite, serpentinite, metaperidotite, and tremolite-talc-chlorite schist. Harper (6) interprets this unit to represent a dismembered ophiolite sequence.

Mining districts occur on both limbs of the South Pass synform. While the South Pass – Atlantic City District occurs along the northwestern limb, the Lewiston District is found on the eastern limb (7). Gold mineralization in the South Pass – Atlantic City District is found chiefly in shear zones in

metagreywacke adjacent to metagabbro sills and dikes. Wall-rock studies of the auriferous shears, show Si and K have been enriched and Ca and Mg have been leached. Mineralogically, these chemical changes are expressed as weak phyllic alteration of the wall rock.   Analyses for native gold from the Diana Mine show high Au/Ag and low Au/Cu ratios (8). The gold analyses and wall-rock alteration are characteristic of a hypothermal vein.

The Lewiston District on the eastern flank of the synform includes strike-trending, metagreywacke – hosted, auriferous shears along the limb of a major fold (9). A few major lodes are localized where the strike shears intersect cross-cutting shears. Wall rocks show distinct chloritic and hematitic alteration as well as weak phyllic alteration.

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EVIDENCE FOR SPREADING IN THE LOWER KAM GROUP OF THE YELLOWKNIFE
GREENSTONE BELT: IMPLICATIONS FOR ARCHEAN BASIN EVOLUTION IN THE SLAVE
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The Yellowknife greenstone belt is located in the southwestern part of
the Slave Structural Province, a Late Archean (2.7-2.5 Ga) granite-
greenstone terrane in the northwestern part of the Canadian Shield.
Supracrustal rocks within this province, collectively referred to as
Yellowknife Supergroup (Henderson, 1970), differ from the supracrustal
successions of the Superior Province and other older Archean terranes by the
absence of komatiites and the high proportion of metasedimentary to
metavolcanic rocks.   The Yellowknife belt was first mapped by Jolliffe
(1942, 1946) on the scale of one inch to one mile, and the gold-producing
area around Yellowknife was remapped on a more detailed scale (1:12,000) by
Henderson and Brown (1966).   As the belt became the best-known example of
the basalt-dominated supracrustal belts in the western Slave Province
(Padgham, 1985), the stratigraphic framework established here (Henderson,
1970), formed the basis for the development of models for Archean basin
evolution (McGlynn and Henderson, 1972; Henderson, 1981).   Under a recent
mapping program of the Geology Division of the Northern Affairs Department
in Yellowknife, detailed mapping was extended, and a 1:10,000 map series for
the entire belt is currently under preparation.   This work resulted in a
number of revisions and refinements in the established stratigraphy
(Helmstaedt and Padgham, 1986) and provides the basis for a reassessment of
current models of greenstone belt evolution in the Slave Province.

The major portion of the Yellowknife greenstone belt is underlain by
the predominantly mafic rocks of the Kam Group which consists of a
northeasterly-striking, homoclinal sequence of flows and tuffs that dip
steeply and face uniformly to the southeast (Fig. 1).   Numerous dikes, sills
and irregular bodies of gabbro and locally anorthosite appear to form an
intergral part of the volcanic sequence.   The Kam Group has been subdivided
into four formations (Fig. 2) with a combined thickness of approximately 11
km.   The lower contact is obscured by the intrusion of a composite batholith
(Western Granodiorite, Fig. 1) that cuts across the strike of the flows.   At
the base of the exposed section, near the northern end of the belt, a narrow
band of felsic volcanic rocks and banded iron- formation is in conformable
contact with overlying pillowed flows above which a mafic extrusive-
intrusive complex is developed (Fig. 2) whose pseudostratigraphy resembles
that of certain Phanerozoic ophiolites.   Near the southwestern end of the
belt, the upper part of the Kam Group (Yellowknife Bay Formation) overlaps a
sequence of older volcanic and sedimentary rocks belonging to the Octopus
Formation (Fig. 1).   In the northern part of the belt, the lower formations
of the Kam Group are truncated by an unconformity beneath conglomerates and
sandstones of the Jackson Lake Formation.   Farther to the south, where the
top of the Kam is preserved locally, it is overlain by calc-alkaline rocks
of the Banting Group that, in turn, are overlain by turbidites of the Walsh
and Burwash Formations.   All rocks of the Yellowknife Supergroup are
deformed and metamorphosed, with metamorphic grade increasing from
greenschist to amphibolite facies towards the granitoid intrusions.   Inspite
of the metamorphic overprint, however, primary structures and intrusive
relationships are well preserved.

The mafic intrusive-extrusive complex of the Chan Formation (Fig. 2)
grades from a lower part, dominated by gabbro, through a multiple dike
complex into massive and pillowed flows with thin beds of interflow
sediments. At the base of the section is a sheet-like body of massive,
medium- to coarse-grained, locally layered gabbro that was intruded into a
sequence of pillowed flows, remnants of which are preserved at three levels.
The upper boundary of this body is a relatively sharp transition into the
dike complex which consists of numerous, fine- to medium-grained metadiabase
dikes and septa and irregular bodies of relatively coarse gabbro between
which screens of pillowed flows can be recognized. The dikes, which are
locally sheeted, show symmetric and asymmetric chilled margins and range in
width from less than one to over 10m. Some dikes grade into pillows,
suggesting that they were intruded close to the seafloor and may have acted
as feeder system to the growing volcanic pile (de Wit and Stern, 1978).
Most of the irregular gabbros are multiple intrusions with abundant chilled
margins and extremely complex contact relationships. Igneous layering is
generally absent at this level, but an up to 100m thick, sheet-like body of
gabbroic anorthosite was recognized (Fig. 2). It is surrounded entirely by
gabbro that has chilled margins against the anorthosite. Though massive and
pillowed flows predominate above the dike complex, sills and irregular
bodies of gabbro, many of them multiple intrusions, are common in the upper
parts of the Chan Formation. The top half of the Kam Group continues to be
dominated by pillowed and massive mafic flows, but contains numerous
intercalations of felsic tuffs and tuffaceous sediments. Some of the flows
and many of the interflow tuffs and sediments are continuous along strike
for more than 10 km and allow stratigraphic correlation across Proterozoic
transcurrent faults (Fig. 1). Synvolcanic mafic intrusions in this part of
the section consist of numerous sills some of which are connected to dike
swarms. The entire section was intruded also by several post-volcanic dike
swarms.

The Yellowknife greenstone belt has been interpreted as the western
margin of an Archean turbidite-filled basin bordered in the east by the
Cameron River and Beaulieu River volcanic belts (Henderson, 1981; Lambert
1982). This model implies that rifting was entirely ensialic and did not
proceed beyond the graben stage. Volcanism is assumed to have been
restricted to the boundary faults, and the basin was floored by a down-
faulted granitic basement. On the other hand, the enormous thickness of
submarine volcanic rocks and the presence of a spreading complex at the base
of the Kam Group suggest that volcanic rocks were much more widespread than
indicated by their present distribution. Rather than resembling volcanic
sequences in intracratonic graben structures, the Kam Group and its tectonic
setting within the Yellowknife greenstone belt have greater affinities to
the Rocas Verdes of southern Chile (deWit and Stern, 1981), Mesozoic
ophiolites, that were formed in an arc-related marginal basin setting. The
similarities of these ophiolites with some Archean volcanic sequences was
previously recognized by Tarney et al. (1976) and served as basis for their
marginal-basin model of greenstone belts. The discovery of a multiple and
sheeted dike complex in the Kam Group confirms that features typical of
Phanerozoic ophiolites are indeed preserved in some greenstone belts and
provides further field evidence in support of such a model.

EVIDENCE FOR SPREADING IN THE LOWER KAM GROUP
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FIGURE CAPTIONS: (1). Geological map of the Yellowknife greenstone belt. Modified from published maps of the Geological Survey of Canada and Northern Affairs Program, Yellowknife. (2). Generalized section of the Kam Group.

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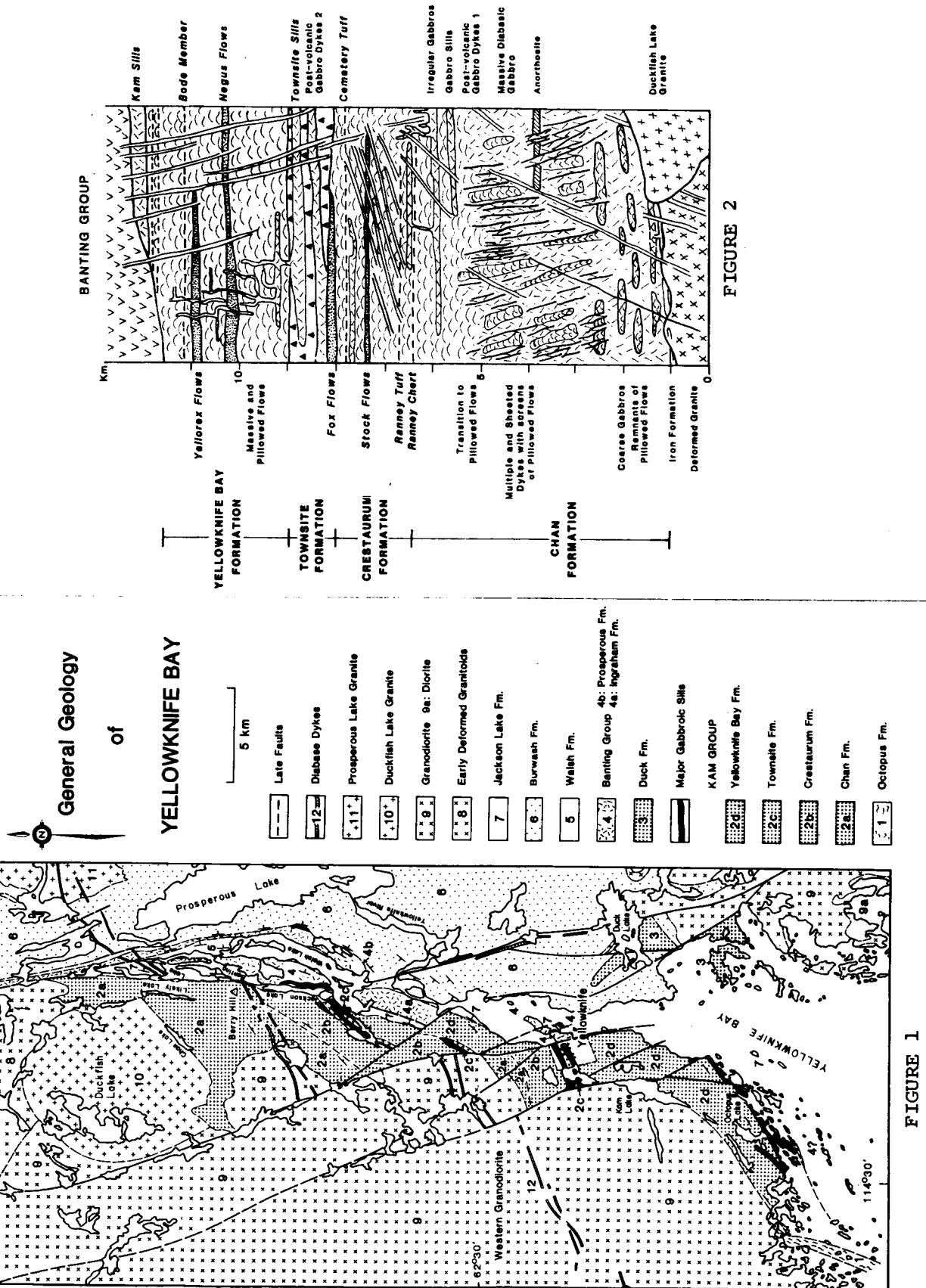
FIGURE 2

FIGURE 1

CRUSTAL ACCRETION IN A 2.7-2.5 Ga "GRANITE-GREENSTONE" TERRANE, SLAVE PROVINCE, NWT: A PROGRADING TRENCH-ARC SYSTEM?

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Progradation of a trench accretionary complex and related magmatic arc provides a simple tectonic model of crustal growth in the Slave Province, NW Canadian Shield. Existing models involving intracontinental rifting do not adequately account for: (1) the paucity of exposed pre-greenstone basement, (2) isotopic evidence for major crust-mantle separation during and after greenstone belt volcanism, (3) absence of coarse clastics intercalated with greenstone volcanics, and (4) the typical stratigraphic sequence of submarine volcanics (tholeiitic and calc-alkaline), veneered by pelagic sediments (chert, iron-formation, carbonate, graphitic pelite), and overlain by turbidites of volcanic+plutonic provenance (whereas in rifts subsidence and clastic sedimentation precede volcanism as the lithosphere progressively attenuates). In the proposed model, the greenstone belts are seen as erosional remnants of a formerly continuous accretionary complex of juxtaposed island arcs and other crustal bathymetric highs, delaminated from subducting oceanic lithosphere and overlain by trench turbidites. Granitoids coeval with greenstone volcanics and fragments of older basement were accreted as arc roots. Subsequently, the foreshortened accretionary complex was extensively intruded by crust and mantle derived plutons of the prograding autochthonous magmatic arc, volcanic levels of which are eroded away. This major plutonism, typically 40-100 m.y. younger than greenstone volcanism, was accompanied by high-T low-P metamorphism of the accretionary complex and provided, in addition to cannibalization of the accretionary complex, a source for the diachronous trench turbidites. Differences between the Slave Province and other Archean granite-greenstone terranes are explained in the model as an accretionary complex dominated by arcs over other types of bathymetric highs (seamounts chains, fracture zones, oceanic plateaus) and a trench kept filled by turbidites, perhaps due to nearby collisional orogeny. The model predicts systematic regional variations in the ages of greenstone volcanism, turbidite sedimentation and autochthonous plutonism. It also predicts that Sm-Nd studies of the autochthonous plutons will yield model ages for bulk crust-mantle separation younger than greenstone belt volcanism, whereas intracontinental rift models predict the opposite.

ARCHAEAN WRENCH-FAULT TECTONICS IN THE ABITIBI
GREENSTONE BELT OF CANADA; C. Hubert and J.N. Ludden, Department
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Structural studies in the southern Abitibi Belt of the
Superior Province have revealed a "dynamic" tectonic style asso-
ciated with wrench-fault systems (1). The fundamental features
of this tectonic regime are the following:

 i) the formation of "lozenge-shaped" blocks of diverse
terranes such as: orthogneissic basement; ultramafic lavas and
associated sediments; tholeiite plateaux and associated sill
complexes; bimodal basalt-andesite - rhyolite volcanic complexes.

 ii) all blocks are bounded either by fault - zones or
by highly strained zones of ductile deformation, and there is a
pronounced gradient in degree of deformation from well preserved
cores to highly deformed and sometimes mylonitized margins;

 iii) sedimentary accumulations occur along the margins
of the blocks in a series of narrow basins bounded by shear -
zones;

The deformation history is summarized below and shown in
simplified form in Figure 1.

The first stage of deformation was simple shearing asso-
ciated with WSW - ESE sinistral wrench faulting which resulted in
NW - SE fold traces, transected schitosities and insignificant
volume changes. Progressive deformation affected blocks of ter-
rane in a tectonic regime in which volcanism, shearing, deforma-
tion and uplift and erosion were synchronous. Terranes composed
of different lithologies were juxtaposed and turbidite accumula-
tions were formed in elongate basins overlying the fault-zones.

The second stage of deformation was a N - S compression
which resulted in the development of highly strained E - W thrust
-shears, fold traces and schistosity and a pressure-solution
cleavage. It also generated NE - SW and NW - SE complimentary
faults defining " S " and " Z " sigmoidal forms and was accompa-
nied by synchronous turbidite accumulation.

The superposition of the second-stage E - W shears on the
first-stage WNW - ESE shears resulted in the formation of "lozen-
ge-shaped" fault-bounded blocks of terrane. These are evident in
the simplified geological compilation of the southern Abitibi
Belt shown in Figure 2.

U - Pb zircon ages, compiled in Ludden et al., (2)
indicate that the volcanic accumulations in the Porcupine, Rouyn-
Noranda and Val D'Or areas of the southern Abitibi Belt define an
axis of volcanism of tholeiitic lineage that was at its peak at
approximately 2700 m.y.. These volcanic rocks superimpose an
older volcano-plutonic terrane which is characterized in the NE -
Abitibi belt and can be correlated towards the SW across the
Kapuskasing front to the Wawa subprovince (2,3). This axis of
volcanism is approximately 2850 - 2720 m.y. in age and is domi-
nated by calcalkaline volcanic and plutonic rocks.

ARCHAEAN WRENCH-FAULT TECTONICS
C. Hubert and J.N. Ludden

A tectonic model is proposed in which the southern Abitibi Belt formed in a series of rift basins which dissected an earlier formed volcanic arc. Comparisons can be made with Phanerozoic areas such as, the Hokuroko basin of Japan, the Taupo volcanic zone of New Zealand and the Sumatra and Nicaragua volcanic arcs. In addition the identification of the major shear-zones makes it possible to speculate that the southern Abitibi Belt comprises a collage of blocks of terrane which have been accreted against a more stable continental margin or micro-continent. If this interpretation is correct, analogies can be made with the SW margin of the U.S.A. in which recently formed blocks of volcanic terrane are being accreted against the western margin of the U.S.A..

FIGURE 1 : Stages of Deformation of the Southern Abitibi Belt.

STAGE 1

STAGE 2

ARCHAEAN WRENCH-FAULT TECTONICS
C. Hubert and J.N. Ludden
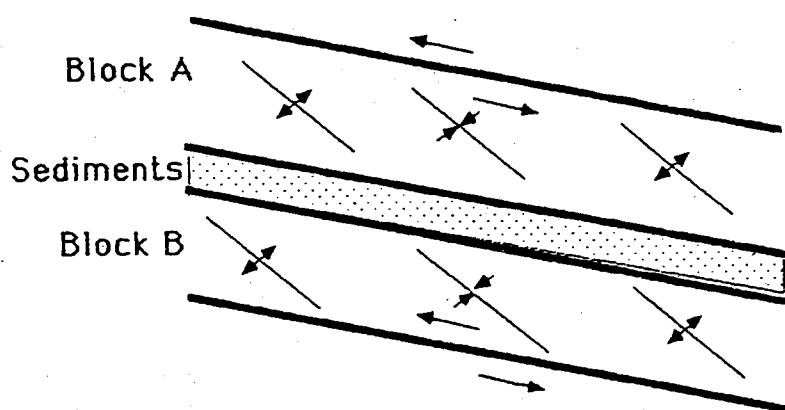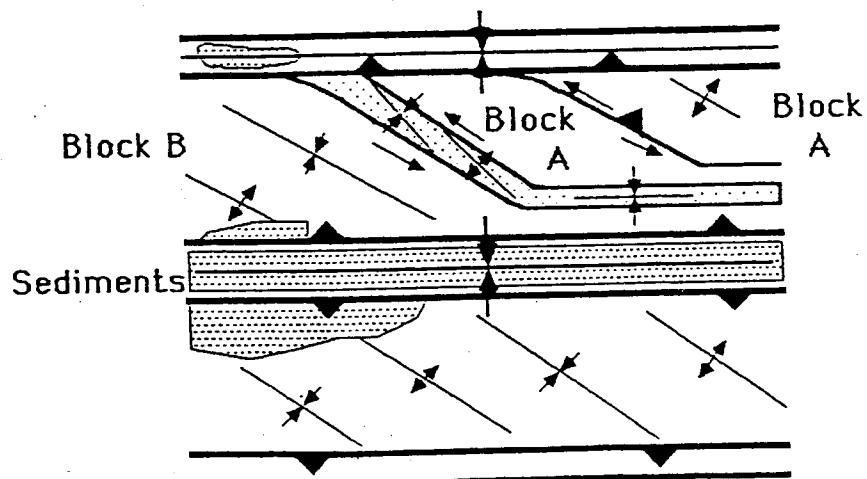
FIGURE 2 : Schematic representation of " lozenge - shaped " blocks of terrane bounded by shear-zones and thrust-shears in the Southern Abitibi belt

INTRUSIVE ROCKS

Qz-MONZONITE-GRANODIORITE

TONALITE-TRONDJHEMITE

GNEISS

VOLCANIC ROCKS

| V V | FELSIC VOLCANIC PREDOMINANT

| O O | MAFIC VOLCANIC PREDOMINANT

ULTRAMAFIC

SEDIMENTARY ROCKS

LATE SEDIMENTARY SERIES

SEDIMENTARY WEDGE

SYMBOLS

SHEAR ZONE
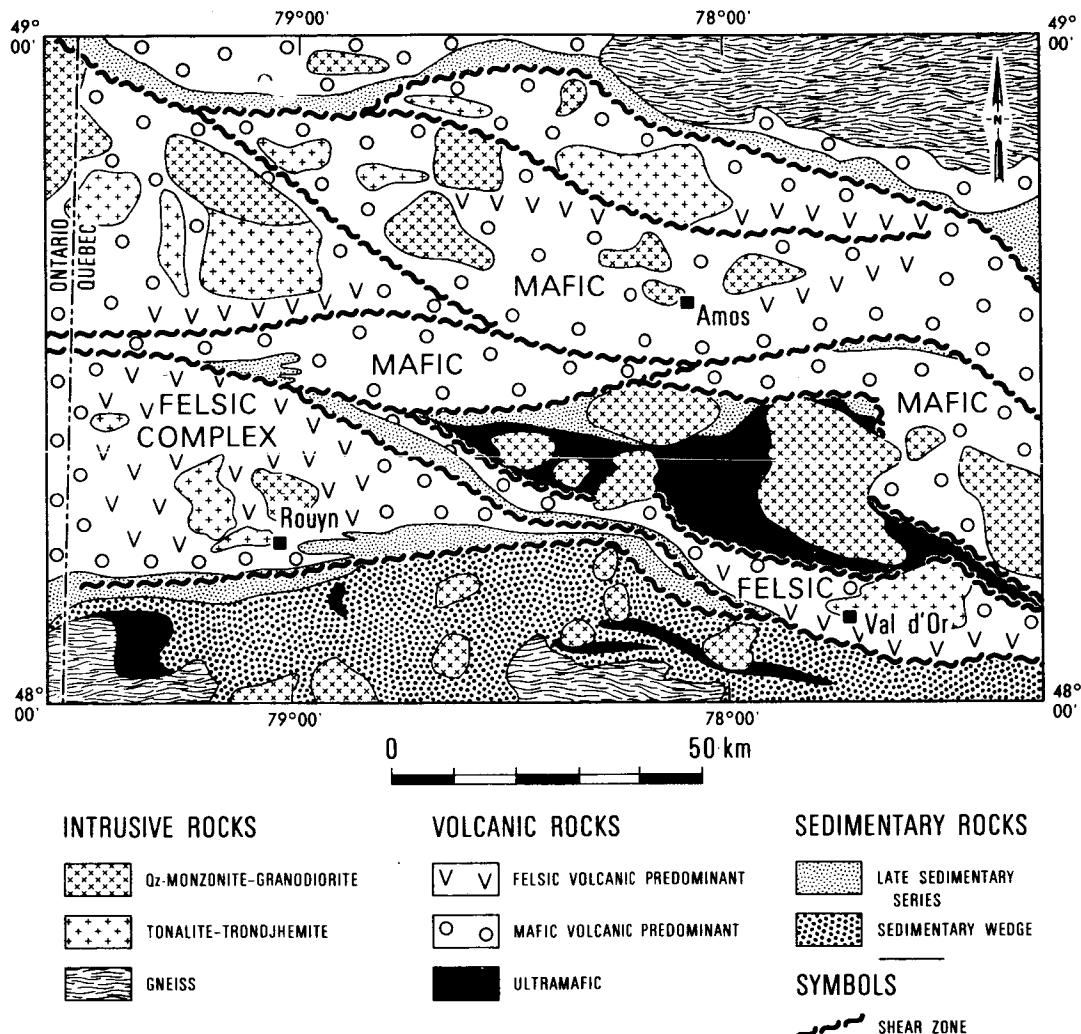
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TRANSPRESSION AS THE MAIN DEFORMATIONAL EVENT IN AN ARCHEAN GREENSTONE BELT, NORTHEASTERN MINNESOTA; P.J. Hudleston and D. Schultz-Ela, Department of Geology and Geophysics, University of Minnesota, Minneapolis, MN 55455; R.L. Bauer, Department of Geology, University of Missouri, Columbia, MO 65211; D.L. Southwick, Minnesota Geological Survey, 2642 University Ave., St. Paul, MN 55114.

Deformed and metamorphosed sedimentary and volcanic rocks of the Vermilion district constitute an Archean greenstone belt trending east-west between higher grade rocks of the Vermilion Granitic Complex to the north and the Giants Range batholith to the south (Fig. 1). Metamorphic grade is low throughout, being lowest in the center of the belt (chlorite zone of the greenschist facies) (1). All the measured strain, a cleavage or schistosity, and a mineral lineation in this belt are attributed to the 'main' phase of deformation (D_2) (2) that followed an earlier nappe-forming event (D_1) (3,4), which left little evidence of penetrative fabric (2).
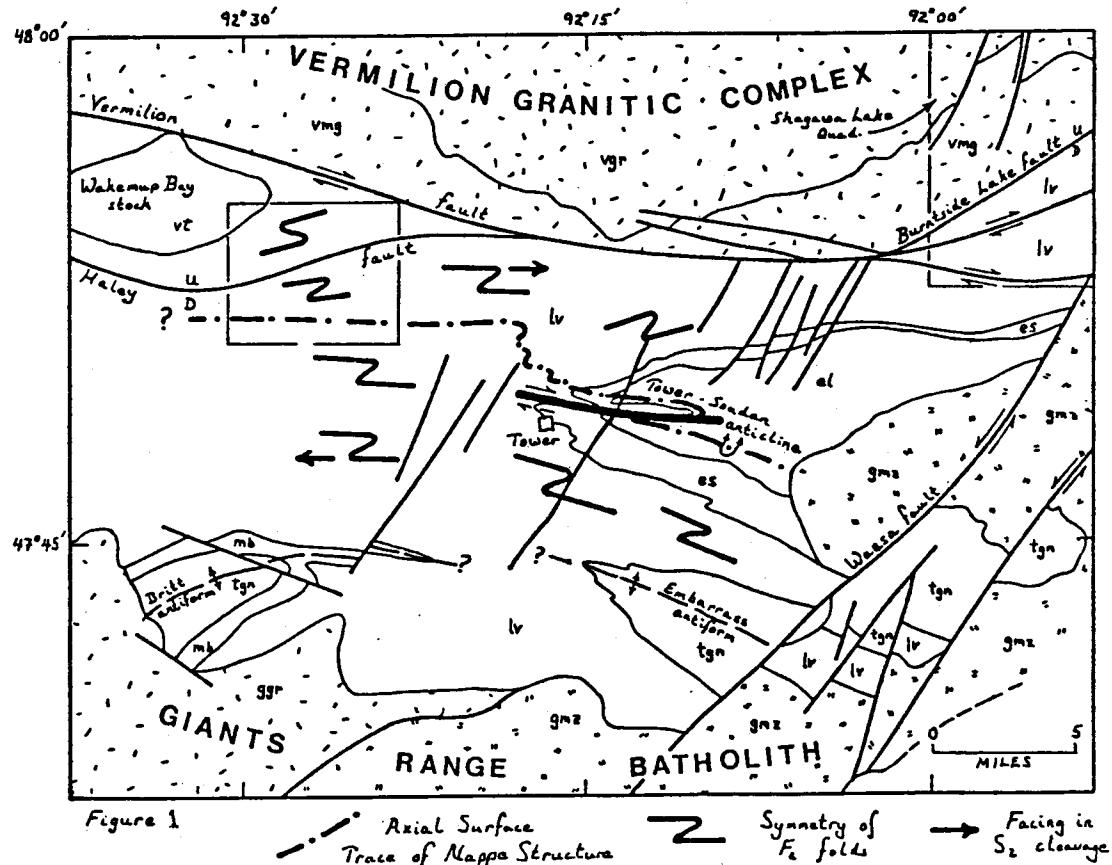
Figure 1. Simplified structural map of the Vermilion greenstone-granite terrane. Rock units are: vgr = granite, chiefly the Lac La Croix; vmg = biotite schist, paragneiss, and migmatite; vt = tonalite; lv = Lake Vermilion Formation plus closely associated rocks of the Upper Member of the Ely Greenstone; es = Soudan Iron-formation Member of the Ely Greenstone; el = Lower Member of the Ely Greenstone; mb = unnamed metabasalt; tgn = tonalite gneiss; ggr = granite; gmz = granite and monzonite.

Previous work assumed that the D_2 deformation resulted from north-south compression across the district, presumably related to diapiric intrusion of the batholithic bodies to the north and south (1). A number of lines of evidence now lead us to believe that a significant component of this defor-

mation resulted from dextral shear across the whole region. Thus the Vermilion fault, a late-stage largely strike-slip structure and one of several faults (1) that bound the Vermilion district to the north, may simply be the latest, most brittle expression of a shear regime that was much more widespread in space and time. Features that are indicative of shear include ductile shear zones with sigmoidal foliation patterns, highly schistose zones with the development of shear bands, feldspar clasts or pyrite cubes with asymmetric pressure shadows, and the fact that the asymmetry of the F_2 folds is predominantly Z for at least 15 km south of the Vermilion fault.

The presence of a large component of simple shear may help explain additional structural features in a simpler way than otherwise possible. Just south of the Vermilion fault the cleavage locally becomes folded and a new cleavage develops in a similar orientation to the old cleavage away from the folds. Rather than interpreting this as evidence for an additional episode of deformation, we consider it to be due to a single process of continuous shear: a foliation develops and after a large strain local perturbations result in folding of the old foliation and the development of a new one axial planar to the folds.

The same type of perturbation can lead to the juxtaposition of ENE-trending zones of constrictional and flattening strains (5) (Figs. 2 and 3), a distinctive feature of the rocks of the Vermilion district otherwise hard to account for. The maximum extension directions (X) of all samples showing constrictional strain, plunge east at angles between 30 and 65°. X in samples showing flattening strain plunges east or west, but near the Vermilion fault all plunges are west or more steeply east than they are in constrictional samples. The maximum shortening direction (Z) plunges consistently less than 25° to the north or south.

The strain variations require a model which can satisfy compatibility constraints and space considerations. The area of consistent constrictional strains in the south may represent one regional component of the strain. Spatial correspondence of flattening strains with the Vermilion fault
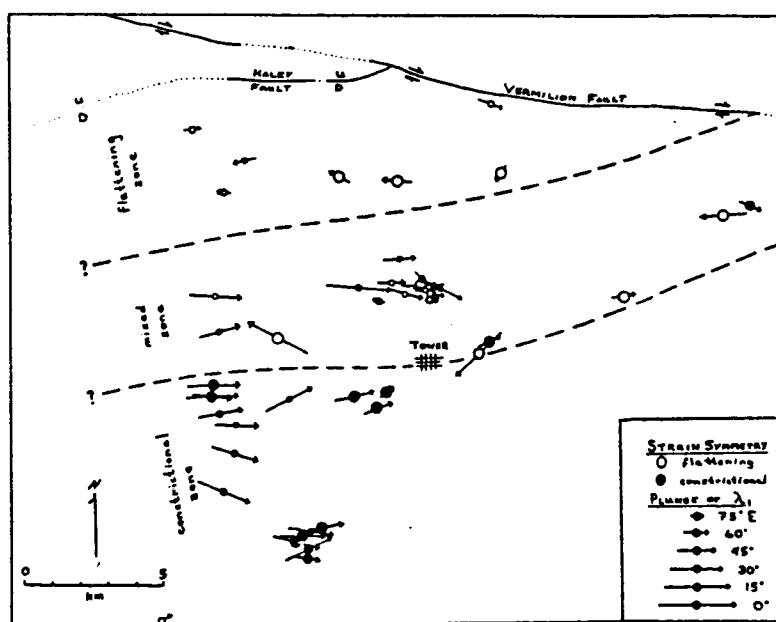
Figure 2. Strain symmetry in metasedimentary and metavolcanic rocks of the Vermilion district. Each dot or circle represents one data point; the size of the dot or circle gives an indication of the reliability of the data.
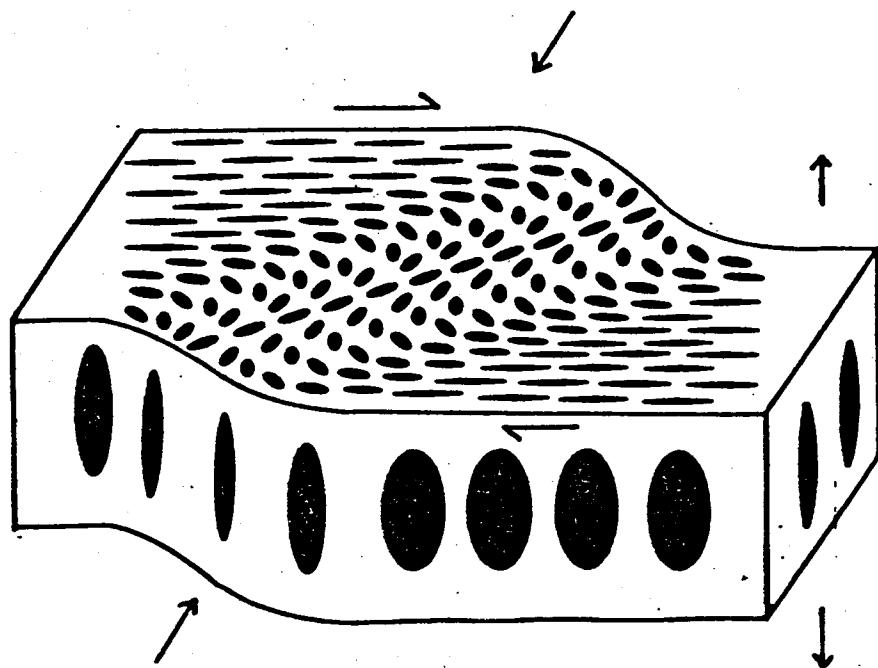
Figure 3. One method for producing bands of juxtaposed flattening and
constrictional strains. Transpression produces flattening strains with
vertical extension, modified in perturbed zone to become constrictional.

suggests that a simple shear component was added in that area. A modified
model of transpression may explain how E-plunging X axes are reoriented to
become W-plunging by a concomitant inhomogeneous progressive simple shear.
Less than vertical plunge of the X axes may necessitate some component of
oblique motion on the fault.

In a general way the strain patterns observed in the Vermilion district
can be reasonably explained by a history of N-S shortening accompanied by
inhomogeneous dextral simple shear. The variations of strain may be a con-
sequence of variations in the relative intensities of shortening and shear,
large perturbations of the shear, or the influences of other structures.
There may be an analogy with the strain partitioning that occurs in small
scale ductile shear zones at large strains.

For transpression to have occurred, the Vermilion district would have to
have been a region of relatively soft lithosphere caught between two more
rigid (either thicker or cooler) blocks to the north and south. We do not
yet know to what extent the high-grade terranes to the north and south were
also affected by transpressional deformation and therefore the configuration
of the more rigid blocks.

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A CONTINUOUS RECORD OF TECTONIC EVOLUTION FROM 3.5 Ga TO 2.6 Ga IN SWAZILAND AND NORTHERN NATAL

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The $\sim$ 3.5 Ga-old bimodal suite underlying an extensive area in southwestern Swaziland comprises the oldest-dated sialic rocks in the Kaapvaal structural province(1). The suite consists of leucocratic, layered tonalitic-trondhjemitic gneisses and amphibolites characterized by the effects of repeated high strains(2). This suite is considered to represent a sialic basement on which metavolcanic and metasedimentary rocks, now preserved as scattered 'greenstone' remnants, accumulated. Direct evidence to confirm this temporal relationship is lacking, but structural data from the Dwalile, Assegaai and Commondale areas indicate that (i) the bimodal gneisses experienced a complex structural history prior to the first recognizable deformation in the supracrustal rocks (i.e. D1 in the supracrustals is equivalent to Dn + 1 in the gneisses) and (ii) scattered remnants of the Dwalile rocks infolded with the bimodal suite structurally overlie the gneisses and are preserved in synformal keels (2)(3). Significant proportions of metaquartzites and metapelites are present in the Assegaai 'greenstone' sequence, the presence of which implies the existence of felsic crust in the source area from which these sediments were derived, a conclusion that is consistent with the structural data.

Ultramafic and pillowed mafic rocks of komatiitic and tholeiitic affinity are present in all four 'greenstone' remnants, but each contains distinctive lithologies. The Assegaai sequence is characterized by the abundance of clastic and chemical sediments that are a minor component of the Commondale and Nondweni remnants. In the former there is a prominent sub-volcanic intrusion composed of multiple layers of massive serpentinite (in which relict cumulate olivine is present locally) alternating with spinifex-textured (olivine and pyroxene) layers. There is a consistent relationship in the thicknesses of the individual layers, i.e. where the serpentinite layers range from 10 to 40 m in thickness the spinifex-textured layers are 1 to 3 m thick. At Nondweni the sequence is dominated by pillowed tholeiites interlayered with high-magnesium basalts and basaltic komatiites (up to 22% MgO). The latter show well developed pyroxene spinifex but peridotite komatiites and units with olivine spinifex are entirely absent. Silicification of the volcanics considered to be contemporaneous with extrusion is not uncommon. Within the volcanic sequence are numerous graded air-fall tuffs and flows of rhyolite compositions. A zone with biogenic or stromatolitic structures is also preserved.

These subtle lithologic differences may reflect different levels of exposure and/or ages of accumulation. The Nondweni greenstones show a consistent northwesterly younging direction in rocks which are not highly strained and which are separated by poorly exposed areas of high strain, suggestive of tectonic interslicing. In contrast the Assegaai and Commondale rocks show evidence of early reclined folds, which may be a reflection of deeper infolding. Preliminary geochronologic data indicate

that the Dwalile 'greenstones' are of similar age to the Barberton greenstones(1). Pb-Pb isotopic data from a single komatiitic flow at Nondweni define an age of 3.15 Ga that is consistent with an Rb-Sr age of $\sim$3.1 Ga for an associated rhyolite(4). However, Sm-Nd data define an age of 3.6 Ga for komatiitic, tholeiitic and rhyolitic flows. Possible explanations are that either the Pb-Pb and Rb-Sr systems were reset at $\sim$3.1 Ga subsequent to extrusion at $\sim$3.6 Ga, or, on eruption 3.1 Ga ago, the extrusions interacted with $\sim$ 3.5 Ga-old felsic crust leading to a range of initial Nd isotopic compositions of the mafic rocks and the generation of rhyolites by remelting of that crust(4).

Subsequent to the D1 event (Table 1), mantle-derived tonalitic plutons (Tsawela and Braunschweig) and the meta-anorthositic Mponono layered intrusive sheet were emplaced into the bimodal gneisses and Dwalile greenstones. All these rocks were strongly and repeatedly deformed under amphibolite-facies conditions (Table 1).

Sheet-like granitoid batholiths were intruded at $\sim$3.2 and $\sim$3.0 Ga, the locus of emplacement migrating northwards with decreasing age. The $\sim$3.2 Ga-old multiphase sodic granitoid intrusion screens the Assegaai and Commondale greenstone remnants from their underlying gneissic basement. Intrusion occurred in the interval between D1 and D2 in the Assegaai and Commondale areas. A chemically and mineralogically similar granite also intrudes the Nondweni 'greenstones' but neither its age nor structural style have yet been studied.

At a high structural level, a second sheet-like, but more potassic granite, the vast multiphase Lochiel batholith, was intruded at $\sim$3.0 Ga north of Dwalile. Following this period of widespread emplacement of granitic magmas emergence above sea-level of stable continental crust took place. Subaerial weathering of this dominantly granitoid terrane was accompanied in the north by the development of braided stream systems draining southeast off the flank of the NE-trending Lochiel batholith(5) into the Pongola basin(6). Minor contemporaneous volcanism accompanied the fluvial sedimentation and heralded a period of subaerial extrusion of lavas (the 2.94 Ga-old Nsuze Group), that range in composition from basalt to rhyolite and attain a thickness of $\sim$ 8.5 km SE of Piet Retief(7). No ultramafic nor high-MgO flow units are present and the sequence is characterized by the simultaneous extrusion of mafic and acidic lavas. Typically porphyritic andesites are also present.

The Nsuze Group is preserved in a series of inliers in the south where its thickness decreases in part due to truncation by the upper (Mozaan) group of the Pongola Supergroup or by the Palaeozoic Natal Group. Volcanic rocks are less abundant in the southern inliers. Shallow water subtidal and tidal-flat sediments including stromatolitic carbonate sands are prominent in the Wit Mfolozi inlier. A heterolithic unit 1.5 km thick dominated by pyroclastic rocks interlayered with shallow marine sediments forms the base of the Nsuze Group south of Babanango. This unit is truncated towards the east by a 4.0 km thick sequence of tidalite sediments with interlayers of basaltic andesite lavas. Transport directions in the inliers are from the north and northwest.

Sedimentation in the Mozaan group was largely controlled by the interaction of a braided alluvial plain and a macrotidal basin(8). Mozaan sediments are not preserved south of the Wit Mfolozi inlier either as a result of removal by erosion or of non-deposition.

The Mozaan Group is typically deformed into gently dipping, doubly plunging synclinal structures resulting from interference of NW and NE-trending axial traces. Adjacent to the southern margin of the Kaapvaal Province, tight E-trending folds with vertical axial surfaces are dominant reflecting a response to deformation related to the development of the Natal thrust zone at ∿ 1.1 Ga. The Nsuze Group is highly strained adjacent to the Swaziland border apparently related to a 20 km wide belt of NW-trending folds and faults with left-lateral movement within which the dyke-like, mafic Usushwana Intrusive Suite was emplaced at ∿ 2.87 Ga(9).

The significance of the Pongola Supergroup lies in the fact that it demonstrates the co-existence of stable continental crust in southeastern Africa and metastable crustal conditions in southern central Africa dominated by extrusion and intrusion of voluminous komatiitic and tholeiitic magmas.

Emplacement of large volumes of granitic magmas principally into Pongola rocks terminated Archaean evolution. Multiple gneiss domes separated by screens of Mozaan sediments of high metamorphic grade developed in southern Swaziland adjacent to the belt of NW-striking, highly strained Nsuze rocks. Subsequently a thin sheet (300 to 1000 m thick) of potassic granite was emplaced at the unconformity between the Mozaan Group and its gneissic granitoid basement. The final pulses of granite plutonism resulted in the emplacement of sharply transgressive, typically coarse-grained, porphyritic plutons ranging in size from 40 km² to 650 km² about which narrow contact aureoles are developed in the Mozaan sediments. Rb-Sr isotopic data have yielded only whole-rock errorchrons for these rocks(10).

The concentration of post-Pongola granitoids within the core of the Pongola depository suggests that depression of the depositional basin promoted partial melting of the lower crust, which would be consistent with the proposed model for the genesis of the granitic melts based on geochemical data(11). The post-Pongola granites differ in their setting from other Archaean granites in southern Africa(5).

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DISTRIBUTION OF
GREENSTONE REMNANTS
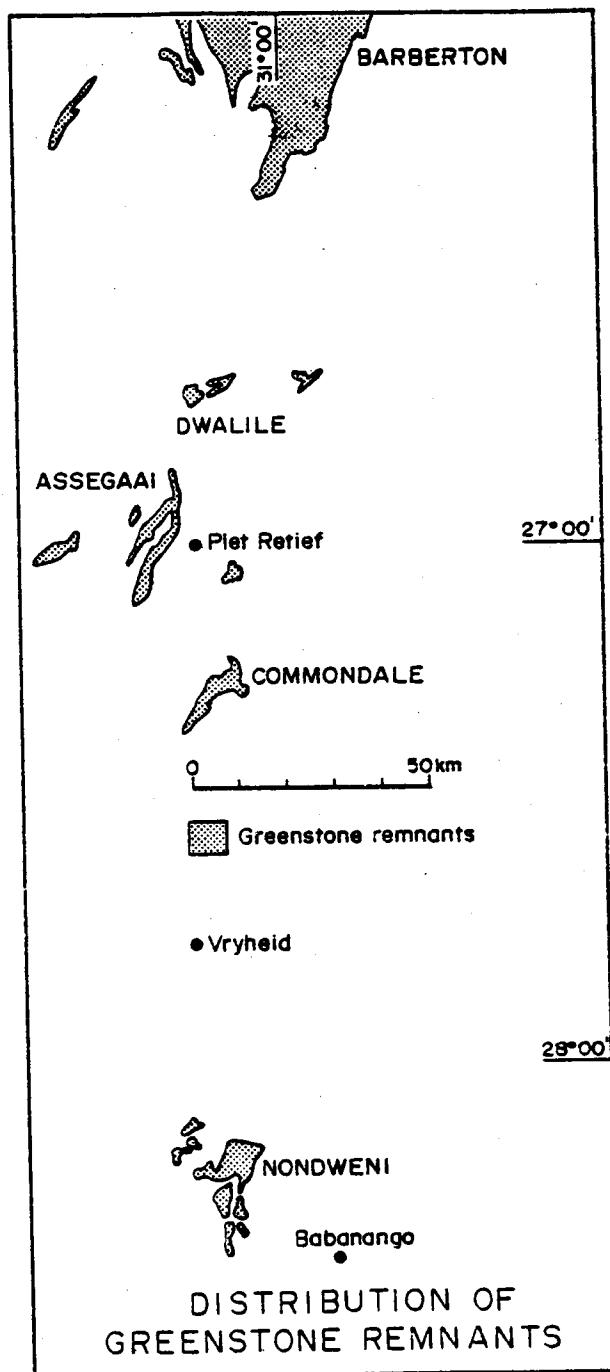
TABLE I   SUMMARY OF GEOLOGICAL AND STRUCTURAL EVENTS PRE-PONGOLA

| SOUTHWESTERN SWAZILAND (Jackson, 1984) EVENTS | | | ASSEGAAI (Talbot, et al., in press) EVENTS | | | COMMONDALE (Smith and Hunter, in prep.) EVENTS | | |
AGE	GEOLOGICAL	STRUCTURAL	AGE	GEOLOGICAL	STRUCTURAL	AGE	GEOLOGICAL	STRUCTURAL
~3.5	Bimodal layered gneisses. Dwalile metamorphic suite (ultramafic, mafic volcanics; minor clastic and chemical sedimentation)	D_1: Flat-lying F folds	?	Bimodal layered gneisses. Assegaai supracrustal suite (ultramafic, mafic volcanics and sub-volcanic intrusions; clastic and chemical sedimentation important)	D_1 (early): Sub-horizontal and listric zones of cataclasis; D_1 (late): Transition to ductile flow; generation of F_1 eastward verging recumbent folds.	?	Bimodal layered gneisses. Commondale supracrustal suite (ultramafic, mafic volcanics and sub-volcanic intrusions; minor clastic and chemical sedimentation)	D_1: generation of recumbent F_1 folds; thrusts parallel to axial surfaces of F_1 folds (?)
~3.3	Intrusion of Tsawela tonalite; Mafic dykes. Intrusion of anorthositic Mponono Intrusive Suite		~3.2 (?)	Intrusion of Anhalt sodic granite sheet		?	Intrusion of Braunschweig tonalite	
?		D_2: Intense F_2 folding on flat-lying XY planes of bulk strain			D_2: F_2 concentric parallel folds with steeply dipping NNE-trending axial surfaces; D_3: Minor dis-harmonic conjugate folds; NE-trending strike-slip faults in meta-sediments; D_4: NW-trending mylonite zones	?	Intrusion of Bazane sodic granite sheet; hornblende granodiorite and related leucotonalite intrusions	D_2: F_2 folds with steeply dipping NNE/NE-trending axial surfaces; D_3: Mylonitic shearing along granite-supracrustal contacts; faults in supracrustals; D_4: Open to tight F_4 folds with steeply dipping NW-trending axial surfaces; D_5: WNW and NW mylonitic shearing
?	Mafic dykes	D_1: local ductile shear zones; D_2: Tight to isoclinal F_4 folds; vertical to gently dipping NE dipping axial surfaces; D_3: Gently dipping ductile shear zones, displacement upwards to NW						
~3.2	Intrusion of quartz monzonite pods		~3.0	Multiple intrusion of Lochiel granite sheet				
~3.0	Multiphase intrusion of Lochiel granite Sheet	D_6: Folding and mylonitic re-foliation of D_5 fabrics near Lochiel granite	~2.96	Pegmatitic granite (?)				

GEOLOGICAL EVOLUTION OF THE PIETERSBURG GREENSTONE BELT, SOUTH AFRICA, AND ASSOCIATED GOLD MINERALIZATION; M.G. Jones, Department of Geology, Imperial College, London and M.J. de Wit, Lunar and Planetary Institute, Houston, Texas.

This poster presents current thoughts based on preliminary field work carried out as part of a Ph.D. project, the aim of which is to integrate the polyphase history of gold mineralization seen in the area with the geochemical and tectonic evolution of the greenstone belt as a whole.

Gold mineralization is found in four distinct regional geological settings:

1. A first phase of gold mineralization was associated with early low grade metamorphism and metasomatism of a 'greenstone basement" sequence of serpentinites (metaperidotites). These are generally intrusive into a series of BIF units, ferruginous shales and cherts. There are also associated extrusive tholeiitic metabasalts and ocellular-bearing komatiitic basalts. The regional hydration which characterizes this early metamorphism resulted in major chemical alteration of the basement and large scale fluid movement, with migration of Fe, and Mg ions, SiO_2 and possibly gold. Early shear zones (possibly represented by a now flat-lying carbonate-fuchsite-gneiss horizon) may have facilitated this fluid movement.

2. The basement sequence is unconformably overlain by a 'cover' of coarse clastic sandstones and conglomerates which contain basement-derived detritus. The conglomerates are often well sorted and graded and may represent coarse turbidites. Placer-type pyrite and BIF clasts, both containing minor gold values, are present in these cover rocks and hence a second period of gold mineralization (reworking) is envisaged.

3. The older rock sequences and gold mineralization above were all affected by a regional deformation event and it is the associated structural traps which contain the most significant gold occurrences seen in this greenstone belt. A well developed upright cleavage with a predominantly NE-SW strike and three major composite shear zones (each containing a number of tectonic breaks) are the main manifestations of this deformation. Strain analysis in the shear zones has been carried out using ocelli from the pillowed komatiitic basalts. The measurements indicate that close to or within the shear zones the finite strain ellipsoid results from a minimum of 50-70% flattening across the cleavage and 100 - 180% extension along the main stretching lineation seam.

Antitaxial and composite extension veins have been recognized. The veins contain fibrous crystals of quartz and calcite which plunge parallel to the stretching lineation (as defined by stretched conglomerate and breccia clasts lying in the cleavage plane). The veins are thus syn-kinematic with this main deformation event. The orientation of the quartz fibres is parallel to the incremental extension growth direction of the dilational veins and so the stretching lineation is parallel to the kinematic movement direction (approx. NW-SE when rotated to the horizontal). The veins are formed by the crack-seal fibrous growth mechanism and semi-quantitative strain analysis indicates clearly that the incremental strain ellipse (in the X-Y plane) did not change orientation significantly during the deformation event.

Field evidence indicates that the shear zones were thrusts (SE over NW) with both vertical and lateral components of movement (Fig. 1). One of the shear zones, the Synmansdrift shear zone is marked by an unusual chaotic breccia which

consists of white and brown-red banded chert and BIF clasts, identical to the BIF from the basement, set in a red ferruginous shale-like matrix. The clast content and size vary abruptly both across and along strike and there is no well-defined bedding (Fig. 2). The lithology can be traced for 7 km along strike and may be up to 100m in thickness but its upper (southern) boundary is ill-defined as it grades over tens of metres into conformably overlying but often highly deformed red shales and sandstones. Hence the upper contact appears to be sedimentary although this has yet to be confirmed by lithogeochemistry. The lower contact is clearly tectonic and an L-S tectonite fabric is well-developed. As well as small clasts, the lower half of the breccia also contains extremely large (up to 100m long x 20m wide) BIF fragments within the lithology. The edges of these larger clasts can be clearly seen to be tectonically ground-up by a 'spalding-off' process which produces the smaller, often euhedral, breccia clasts.

As a whole the unit constitutes a tectono-sedimentary melange which is envisaged to have formed as a sedimentary wedge above a low dipping shear zone (thrust) during horizontal shortening across the region. Large scale movement of Fe ions, $SiO_2 \pm Au$ occurred (Fig. 1).

Gold mineralization is found in quartz $\pm$ tourmaline veins associated with various structural traps e.g. fold hinges and minor shear planes including ultracataclasites. In the vicinity of these traps pressure solution and metamorphic segregation features are common which indicate fluid movement and possible gold mobility from the deformed sediments (and possibly the basement rocks) into the traps. This fluid migration may have occurred early with respect to the defomation with the resultant veins being subsequently slightly deformed and tectonically displaced.

4. A later porphyroblastic overprint of gold-bearing arsenopyrite is seen locally within the shear zones as well as porphyroblasts of ephesite (a lithium-bearing brittle mica) and andalusite. These features seem to indicate a later period of gold mineralization and 'static' metamorphism probably related to granitic intrusions which provided a heated source (and possibly fluids) for element mobility and mineralization within the already deformed volcano-sedimentary pile.

Fig. 1. Schematic representation of the inferred structural evolution (A-D) across area during Dm(north-central part of the Pietersburg greenstone belt). In this scheme, an older (~3.4-3.5 Ga) mafic-ultramafic/BIF basement is deformed and syntectonically overlain by coarse clastics, some of which are demonstrably derived from the "greenstone basement".

Fig. 2. Different aspects of the tectono-sedimentary melange. Upper photo shows large clast in finer matrix. Mega clasts may be up to 10-20m in length. Lower photo shows large variations in aspect ratios of clasts. Cleavage is parallel to the hammer-head.

SECTIONS

N — S

A GREENSTONE — UPLIFT AND EROSION
BIF

B CONGLOMERATES
MELANGE

C

D
~75%
KUSCHKE BIF
HOLLANDSDRIFT BIF
REGIONAL CLEAVAGE Sm

1 HOLLANDSDRIFT SHEAR ZONE
2 SNYMANSDRIFT MELANGE - SHEAR ZONE
3 KUSCHKE SHEAR ZONE

A
BIF
SE
NW

B
MELANGE
BIF
CRUSTAL THICKENING
FLUIDS & AU

C
CONGLOMERATES (with placer gold)
MELANGE
HOLLANDSDRIFT
KUSCHKE
MELANGE

FIG.1

C = Clast ; M = Matrix
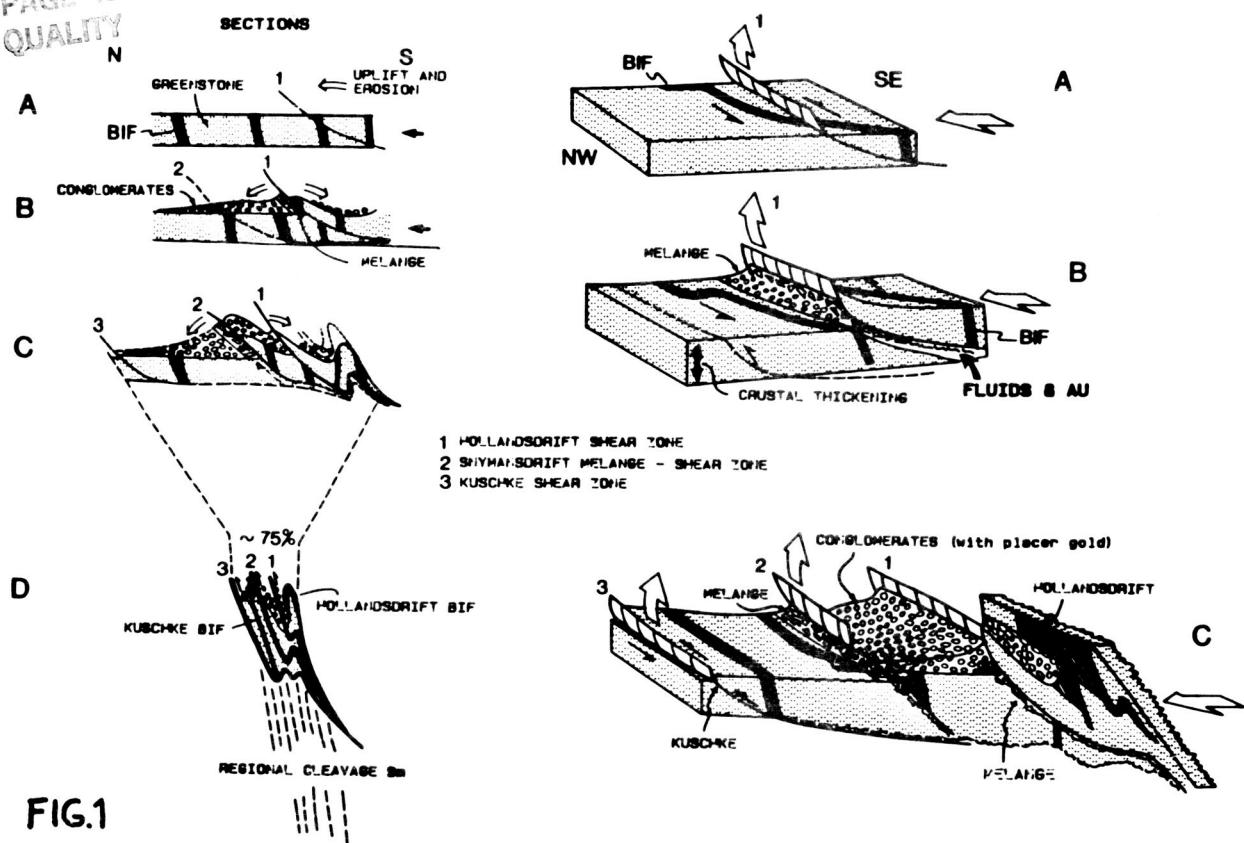
FIG.2

ARE GREENSTONE BELTS IN THE SLAVE PROVINCE, N.W.T. ALLOCHTHONOUS?

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Many tectonic models for the Slave Province, N.W.T., Canada, and for Archean granite - greenstone terranes in general, are implicitly dependent on the assumption that greenstone belt lithologies rest unconformably upon older gneissic basement. Other models require originally large separations between gneissic terranes and greenstone belts. A key question relating to the tectonics of greenstone belts is therefore the original spatial relationship between the volcanic assemblages and contended-basement gneisses, and how this relationship has been modified by subsequent deformation. From the Slave Province, unconformities have been reported from the base of the Cameron River Greenstone Belt northeast of Yellowknife, and from the Point Lake area to the north (Figure 1 and refs. 1,2). What remains unclear in these examples is the significance of the so-called "later faulting" of the greenstone - gneiss contacts. Does the angular discordance between greenstones and gneissic foliation in the Cameron River example really represent an unconformity, or could it be better- interpreted as a consequence of the juxtaposition of two once widely separated terranes? Where unconformities between gneisses and overlying sediments are indisputable, such as at Point Lake (and also in the Belingwe greenstone belt in South Africa), the significance of faults which occur below the base of the volcanic succession also needs to be evaluated. As part of an on-going investigation aimed at answering these and other questions, I mapped the extremely well-exposed contact region between the Cameron River Greenstone Belt and the Sleepy Dragon Metamorphic Complex in the vicinity of Webb Lake and Sleepy Dragon Lake during the summer of 1985, extending the database of earlier workers (3,4,5,6,7).

The greenstone belt was found to consist predominantly of mafic pillowed to massive flows and numerous dike complexes. At the preserved base of the belt these dikes locally retain a sheeted aspect and display one-way chilling. Subordinate amounts of pyroclastic rocks and volcanic breccias are also present. Rocks of the Sleepy Dragon Metamorphic Complex are highly variable, and include both meta-sedimentary and meta-igneous gneisses, along with numerous mylonite zones (4,8,9). Older gneisses and mylonites are intruded by several younger phases of mafic to silicic plutonic rocks which show different intensities of deformation.

The contact between the Cameron River Greenstone Belt and the Sleepy Dragon Metamorphic Complex was found to be a half-kilometer wide zone of very complex structure. All rocks within this high-strain zone have a strong steeply plunging stretching lineation, although rocks from throughout the area also have a less-intense vertically plunging lineation. Transposed layering and intensely folded quartz segregations are common in this zone; sheath folds with vertically plunging hinges are present in some localities, indicating very high shear strains. Macroscopic sense-of-shear indicators are not abundant but generally suggest that the Cameron River Belt was thrust over the Sleepy Dragon Complex. Supporting microscopic work

is currently underway. In one area north of Webb Lake some slivers of Sleepy Dragon-type gneisses are intercalated with phyllonites of the major high-strain zone. In this locality, ductile shear zones cutting the Sleepy Dragon gneisses are oriented in a way that is suggestive of mylonitization contemporaneous with thrusting of the greenstones over the gneisses. Further mapping will reveal the lateral extent of this structural juxtaposition, but it is apparently the first-documented example of Archean basement-involved thrusting of greenstones over gneiss in the Slave Province.

Pillow lavas of the Cameron River Belt immediately adjacent to the basal high-strain zone have aspect ratios exceeding 3:1:1/3. In an area extending northward from Sleepy Dragon Lake these lavas are overturned in an isoclinal fold as shown by locally-consistent younging directions. The axial trace of this fold is parallel to the contact zone, and the fold's geometry is consistent with formation during thrusting of the Cameron River Belt over the Sleepy Dragon Complex. Preliminary mapping of the greenstone belt in the Webb Lake area has revealed the presence of a few other subparallel shear zones containing structures similar to those just described; a common origin is tentatively inferred pending more detailed mapping.

Interpreting the structures within the Sleepy Dragon Metamorphic Complex is difficult because of the complex deformation history of this terrane. The only structure which, at this point, can unambiguously be related to movement along the contact with the Cameron River Belt is a foliation which trends parallel to and increases in intensity towards the contact zone. The foliation cuts earlier structures including folded gneissic and mylonitic foliations; earlier foliations are folded about this later one (4). The fact that this late foliation is cut by some plutonic bodies suggests that a minimum age may be placed on the thrusting and emplacement of the Cameron River Greenstone Belt over the Sleepy Dragon Metamorphic Complex.

Numerous mafic dikes are present both at the base of the Cameron River Belt and within the Sleepy Dragon Complex near it's contact with the greenstone belt (7). The textures and xenolith contents of the dikes in the Sleepy Dragon Complex appear to be generally different from the dikes in the greenstone belt. Deformational and metamorphic fabrics in the dikes of the Sleepy Dragon Metamorphic Complex suggest that they are of at least two, and probably three generations, while only two distinct generations of dikes are recognized from the Cameron River Greenstone Belt. Pending further field and laboratory work it is tentatively suggested that (a) the first two generations of dikes in the Sleepy Dragon Complex are not directly related to any dikes in the greenstone belt, (b) the earliest generation of (locally sheeted) dikes in the greenstone belt is not present in the basement complex, and (c) only the latest, relatively undeformed dikes are correlatable between the two terranes.

Although an unconformable relationship has been reported to exist between the Cameron River volcanic belt and the Sleepy Dragon Complex, I have not yet observed it. All data collected to-date indicates that the greenstone belt is allochthonous. Structures at and near the base of the greenstone belt suggest that it has been imbricated and thrust over the Sleepy Dragon Complex, incorporating
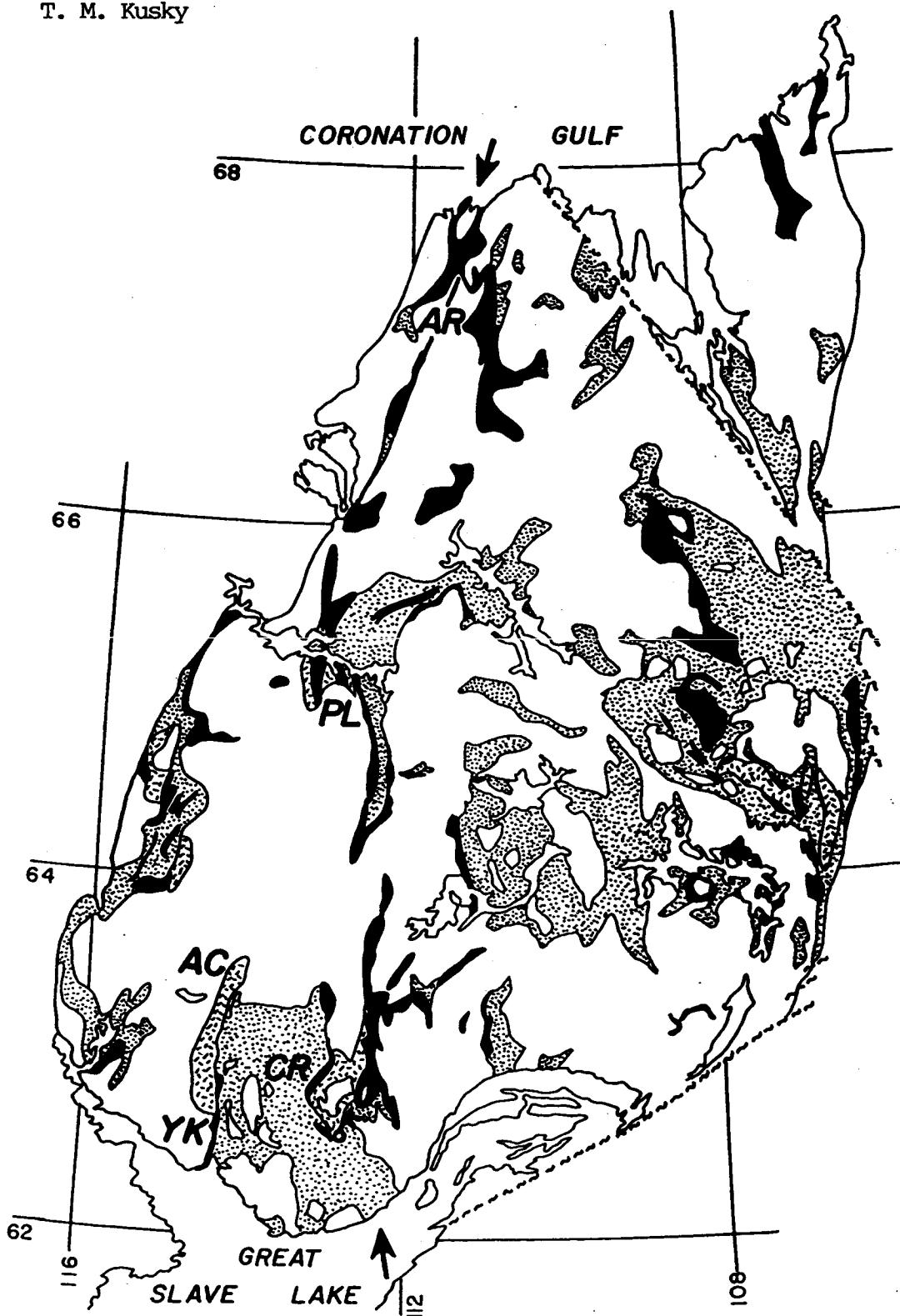
Figure 1. Map of the Slave Province, N.W.T., showing the
distribution of "greenstone" belts (black), gneiss complexes (cross
hatch), and graywacke sediments (stippled). Symbols: YK=Yellowknife
Greenstone Belt, CR=Cameron River Greenstone Belt, AC=Anton Gneiss
Complex, PL=Point Lake, AR=Anialik Greenstone Belt. Arrows denote
proposed suture which separates the eastern and western greenstone
terranes. Map after 3 and 12.

slivers of gneiss in the basal high strain zone in the process. A crude order-of-magnitude estimate of the amount of displacement needed to explain the observed deformation between the Cameron River volcanics and the Sleepy Dragon gneisses is several, if not tens, of kilometers. Actual displacements are likely to be significantly higher.

The present poor constraints on the sense and magnitude of displacement along the fault complex separating the Cameron River Belt and the Sleepy Dragon Complex, as well as a paucity of similar data from other preserved greenstone/gneiss contacts in the Slave Province (Point Lake, Anton Complex; Figure 1), allows a set of "permissible" tectonic models for the Slave Craton to be formulated at this time. It is tentatively proposed that the Sleepy Dragon Complex is a preserved remnant of an Andean arc complex. This suggestion is inferred because: (1) the prolonged magmatic and deformational history of this terrane is typical of Andean arc settings; (2) the generally mafic to intermediate volcanic suite is similar to that found in Phanerozoic Andean arc settings and; (3) the composition and intrusive style of plutons is identical to more recent Andean intrusive suites. The preserved extent of this Andean-type arc complex could be defined by the aerial distribution of Sleepy Dragon type gneisses (such as the Anton Complex?), which all seem to occur west of a prominent N-S striking line of "greenstone" belts and major faults that extends from the Great Slave Lake to Anialik River Belt on the Coronation Gulf (Figure 1). Volcanic belts to the east of this line have a much greater abundance of silicic volcanics than belts to the west (12). A preliminary interpretation of this line is that it represents a suture between the eastern and western greenstone terranes. The eastern greenstone belts might represent an amalgamated island arc complex, or a collage related to a migrating arc-trench system (13), which collided with and became sutured to the Sleepy Dragon Andean arc complex. While it is quite possible that no pre-deformational link exists between the Cameron River and the Yellowknife greenstone belts, their similarity is conspicuous. Much data suggests that the Yellowknife and Cameron River "greenstone belts" are back arc basin "ophiolitoids" (10, 11). This model is significantly different from earlier studies which concluded that the basalts were erupted in a continental rift setting (3). The abundant "Burwash Formation" turbidites do not appear typical of continental rift deposits but do strongly resemble more recently deposited syn- to post-orogenic flysch and molasse sequences. An accretionary wedge origin is also possible for this complex sedimentary package (13).

Thus, it seems possible that prior to accretion of the island arc terranes to the east of the N-S striking line of "ophiolitoids", the Sleepy Dragon Andean arc complex experienced a back arc rifting/spreading event which formed the Yellowknife and Cameron River Belts. Some of the Sleepy Dragon gneisses and phyllonites below the Cameron River Belt, and sediments found at the "base" of the Yellowknife Belt (Octopus Formation) could represent a volcanosedimentary sequence deposited during this intra-arc rifting event or, alternatively, they might be precursory olistostromes related to the emplacement of the greenstone belts. Further detailed structural observations are needed to decide between these

hypotheses.

Structural relationships between the Cameron River Greenstone Belt and the Sleepy Dragon Complex are thus compatible with either the emplacement of a back arc basin "ophiolitoid" over the Sleepy Dragon gneisses, or with the emplacement resulting from the closure of a major ocean. Similar structural relationships appear to exist elsewhere in the Slave Province where greenstone/gneiss contacts are preserved, suggesting that all greenstones terranes in the region may be allochthonous. It will be interesting to see how these preliminary models stand up to the test of several more seasons of detailed field work in the Slave Province.

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SYNSEDIMENTARY DEFORMATION AND THRUSTING ON THE EASTERN MARGIN OF THE BARBERTON GREENSTONE BELT, SWAZILAND, Lamb, S.H., Research School of Earth Sciences, Victoria University of Wellington, Private Bag, Wellington, New Zealand.

Mapping on the eastern margin of the 3.6-3.3 Ga Barberton Greenstone Belt, NW Swaziland, has revealed a tectonic complex which is more than 5 km thick (Lamb, 1984a). The area consists of fault bound units made up of three lithological associations. Some of these have been affected by four phases of deformation (D1-D4). Fold structures (F1-F4), foliations (S1-S4), and lineations are associated with the deformation.

The oldest rocks consist of metaigneous rocks (talcose schists, serpentinite, and quartz-chlorite-sericite schists) interleaved with silicified fine grained sediments (cherts). These make up the Onverwacht Group, though deformed (D1) and intruded by meta-ultramafic rocks. Onverwacht Group cherts locally pass conformably into a circa 1.8 km thick sequence of siltstones, shales, BIF, with sandstone and conglomerate layers, forming the Diepgezet Group. The lower part of the Diepgezet Group is interpreted as submarine fan deposits, and can be correlated with sequences in South Africa referred to as both the Moodies and Fig Tree Groups (Lamb and Paris, in prep). The Diepgezet Group is overlain unconformably, with angular discordances of up to 90 degrees, by at least 1.8 km of coarse clastics (Malalotsha Group). These are interpreted as fluvial and marginal marine deposits. In certain localities the Diepgezet Group passes up conformably into the Malalotsha Group through a sequence of coarse sediments which have been left undifferentiated (Mal/Diep Group). Parts of the Malalotsha Group can be correlated with the Moodies Group.

Three pronounced angular unconformities occur within the basal 1000m of the Malalothsha Group. Malalotsha Group sediments are both folded by, as well as unconformably overlying, D2 fold structures which deform the Diepgezet and Onverwacht Groups. Folded fault zones (D1) juxtaposing the Diepgezet and Onverwacht Groups are also unconformably overlain by the Malalotsha Group. Faults associated with the F2 folding (flexural slip faults) offset Malalotsha Group sediments, but are also unconformably overlain by younger Malalotsha Group sandstones and conglomerates. In sequences where the Malalotsha Group is transitional with the Diepgezet Group, a progressive change is observed in the clast content of the sandstones. Chert grain dominated sandstones within the Diepgezet Group pass up into sandstones made up mainly of single crystal quartz grains. Clasts representing all the underlying stratigraphy, as well as parts of the gneissic terrain (potassium poor granitoids) are found in Malalotsha Group conglomerates. Palaeocurrents within the basal Malalotsha Group indicate polymodal sediment transport directions. This, combined with evidence for rapid sediment thickness changes and facies variation, suggest that these sequences were deposited in tectonically controlled (and actively deforming) basins. However the overall tectonic setting is not clear, though the sediments were clearly deposited in a compressional regime.

The sedimentary sequences described above are now found within thrust sheets up to a kilometre thick. These are bounded by thrust faults, subparallel to bedding, which juxtapose different parts of the stratigraphy. One of these thrusts emplaces part of the Onverwacht Group on top of the Malalotsha Group, with a displacement of more than 10 km. The Onverwacht Group here contains a low angle foliation (S2) subparallel to the bounding fault. The thrust faults are considered to be a later expression of the D2 deformation, which is seen as syn-sedimentary deformation structures within the thrust sheets. The D2 deformation caused shortening in northerly and westerly directions.

The thrust sheets and their internal structures have been refolded by tight

kilometre scale north trending folds, which plunge south at 20-40 degrees (F3). The folds contain a pronounced axial planar cleavage defined in places by a muscovite schistosity. The cleavage is most intense near and within marginal granitoids which were probably intruded c. 3.0 Ga (part of the Mpuluzi batholith, Barton 1981). Earlier fold structures have been tightened up, intensifying an axial planar cleavage fabric in F2 folds (S2/S3). The contact with intrusive granitoids on the western margin of the study area (Steynsdorp pluton, which may be c. 3.4 Ga, Barton 1981) contains a pronounced foliation which cuts across intrusive contacts. This is interpreted as an S3 foliation which contains an intersection and/or stretching lineation plunging at 20-40 degrees NE. The apparent domal pattern of foliations in the marginal parts of the Steynsdorp pluton is interpreted as both the result of F3 folding of an earlier foliation (S2) and also the imprint of an S3 foliation. Elongation lineations in sediments within the greenstone belt may be a result of subvertical extension during the D3 shortening (e.g. Jackson and Robertson, 1983).

The above structures have been refolded by heterogeneous southeast trending folds (F4) with the local development of an L4 crenulation lineation.

It has been suggested (Lamb, 1984a,b) that the high level syn-sedimentary D2 deformation and subsequent development of a thrust complex was related to coeval deformation and metamorphism (Jackson, 1984) in the Ancient Gneiss Complex of southern Swaziland. D2 in the study area predates the c. 3.0 Ga Mpuluzi batholith. It is not clear what the relation was between D2 and an early D1 deformation, which occurred during the evolution of the Onverwacht Group rocks (de Wit, 1982; pers. com.). It is likely to be close as a continuous depositional sequence is preserved between the Onverwacht and Malalotsha Groups. The correlation of clastic sequences in the southern part of the greenstone belt with those in the study area, indicates that the D2 deformation was diachronous with variable structural trends. The presence and position of unconformities show that NW-SE shortening (D2b) and the deposition of the Malalotsha Group in the study area post-dates the deposition of the Moodies Group and N-S shortening (D2a) observed in the southwestern part of the greenstone belt (de Wit et al., 1983). It is however not clear to what extent the D2b shortening has reworked and translated structures which formed in D2a. Subsequent D3 deformation (coeval with the intrusion of the Mpuluzi batholith, c.f. Jackson and Robertson, 1983) has had a considerable effect on structures in the study area, continuing the shortening (E-W) on the eastern margin of the greenstone belt.

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THE ROCK COMPONENTS AND STRUCTURES OF ARCHEAN GREENSTONE BELTS:
AN OVERVIEW; Donald R. Lowe and Gary R. Byerly, Department of Geology,
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Much of our understanding of the character and evolution of the
earth's early crust derives from studies of the rocks and structures in
Archean greenstone belts. Our ability to resolve the petrologic, sedimento-
logical, and structural histories of greenstone belts, however, hinges
first on an ability to apply the concepts and procedures of classical
stratigraphy. Unfortunately, early Precambrian greenstone terranes present
particular problems to stratigraphic analysis, some of which we would like
to discuss here. We would also argue that many of the current contro-
versies of greenstone belt petrogenesis, sedimentology, tectonics, and
evolution arise more from our inability to develop a clear stratigraphic
picture of the belts than from ambiguities in its interpretation.

We will here consider four particular stratigraphic problems that
afflict studies of Archean greenstone belts: (a) determination of facing
directions, (b) correlation of lithologic units, (c) identification of
primary lithologies, and (d) discrimination of stratigraphic versus struc-
tural contacts.

(a) Facing Directions: Determination of facing directions in green-
stone belt sequences is often difficult because of the absence of useful
facing indicators throughout great thicknesses of section and because we do
not sufficiently understand the origins of many structures and textures in
Archean sedimentary rock types to be able to use them as facing indicators.
Thick sequences of massive volcanic rocks, banded black and white cherts,
black cherts, and banded iron formation are inevitably rather stingy in
yielding familiar facing indicators whereas thick turbiditic units, layers
of graded accretionary lapilli, and sands containing large-scale cross-
stratification are particularly user-friendly in this regard. Facing
directions in banded cherty units are most readily determined from fluid
escape features, particularly pockets of druzy quartz, which originate as
pockets of trapped fluid, usually directly beneath early-lithified white
chert bands. Geopetal accumulations of debris in cavities, cracks, and at
the bases of early-formed breccias and the preferential development of
stalactitic dripstone in stratiform cavities (the development of both
stalactitic and stalagmitic dripstone is also common, but stalagmites alone
are extremely rare) are also widespread and useful as facing indicators in
cherty successions. In all cases where supporting evidence is available in
adjacent sedimentary units, we have found pillow geometry and drain-out
cavities, where developed, to be reliable facing indicators in tholeiites.

Small-scale cross-laminations, load structures, and individual graded
detrital layers must be approached with caution because nearly identical
features can form facing upward or downward. Pillows, where present in
komatiitic sequences, generally lack useful facing information. The recent
trend to quantify the reliability of facing estimates (e.g. 95% confidence)
is misleading inasmuch as the principal errors in determining facing
directions originate not through statistical ambiguities in the structures
themselves but from their misidentification by the investigator.

(b) Correlation: The correlation of stratigraphic units within poorly
exposed, structurally complex, highly altered Archean terranes represents a
major challenge to unravelling greenstone belt stratigraphy and evolution.
The absence of useful guide fossils and the paucity of unique, recognizable

time markers, such as distinctive ash beds, makes this task difficult
relative to similar studies in Phanerozoic terranes. Recent precise zircon
age dating in the Canadian belts is aiding in resolving gross problems of
stratigraphy, but will do little for detailed correlation.

In the early Archean Barberton and Pilbara belts, we have found a
number of features particularly useful in correlation: (1) lithologically
and texturally distinctive layers of airfall and/or turbiditic accretionary
lapilli, (2) individual airfall ash beds in sequences of orthochemical and
biogenic deposits, (3) airfall spherule layers, (4) distinctive sequences
of non-facies controlled deposits, and (5) rare, facies-related units and
sequences. Least reliable are distinctive successions of environmentally
or petrogenetically controlled lithologies that can be repeated many times
within individual sections as sedimentary environments and magmatic systems
come and go. Even continuous, traceable lithologic units cannot serve as
unambiguous time markers unless there is independent evidence that they are
not diachronous.

(c) Primary Lithologies: Perhaps as much as any other problem, our
inability to decipher primary lithologies has hampered the development of a
clear picture of greenstone belt make-up and evolution. It has long been
recognized that early alteration is pervasive throughout greenstone belts.
This alteration was for many years considered part of the post-accumulation
metamorphic history of these belts. More recently, however, the trend has
been to attribute alteration to relatively high-temperature exhalative to
shallow-subsurface hydrothermal processes (1, 2) or to low-temperature
metasomatism, perhaps related to the circulation of surficial waters
through the rock sequences (3).

Interpretation of the primary MgO contents and petrogenesis of koma-
tiites, role of calc-alkaline and subduction-related volcanism, presence or
absence of volcanic cycles, distribution of felsic lavas, nature of meta-
morphism and metasomatism, provenance of detrital sediments, composition of
early surface waters, and sedimentology of cherty units have all been
stymied to some extent by uncertainties in the composition of the original
sedimentary and volcanic layers. A number of relatively recent studies
have shown clearly that (i) many specific units previously interpreted to
be silicic volcanic rocks are actually silicified mafic to ultramafic lavas
(e.g. 2, 3), (ii) many of the "classic" mafic-to-felsic volcanic cycles are
non-existent (4) although large-scale volcanic cyclicity seems to be widely
developed (5), (iii) calc-alkaline volcanics, as well as komatiites, are
abundant in some belts but poorly represented in others, (iv) some belts
exhibit a more-or-less continuous spectrum of rock compositions from
komatiitic to rhyolitic whereas others are strongly bimodal or trimodal;
(v) evaporitic sediments, especially gypsum, were widespread and abundant
constituents of shallow-water Archean greenstone-belt sedimentary deposits
(6), (vi) relatively few, if any, cherty layers represent primary silica
precipitates (7), and (vii) there may be important lithologic and tectonic
differences between early and late Archean greenstone belts (7).

Many of the remaining ambiguities in the alteration histories of these
rocks originate because most studies of alteration are focused on identify-
ing the role or evaluating the influence of one particular style or setting
of alteration. Clearly, some silicification and carbonatization began
concurrently with deposition and involved essentially surface waters at
surface temperatures. The abundance of cherts in shallow-water sequences
but their paucity in deeper-water units (7) suggests that early post-

depositional fluctuations in water chemistry (e.g. deposition in marine but
early flushing by meteoric waters) may have been an important control on
silicification. Later large-scale recrystallization and replacement almost
certainly occurred both through low-temperature processes, similar to those
affecting modern oceanic crust, as well as during local higher-temperature,
hydrothermal and black-smoker-type metasomatism and mineralization. The
widespread presence of epidote and resetting of isotopic systems, such as
Ar-Ar, clearly argue for still later regional metamorphism, and the locali-
zation of silicification along some joints and fractures indicates continued
alteration under fully post-tectonic and post-metamorphic conditions.
Future studies must provide unambiguous criteria for distinguishing stages
and environments in this prolonged alteration history, many of which may
leave similar mineralogical and textural records.

(d) Stratigraphic vs. Structural Contacts: Greenstone belt sequences
are characteristically highly deformed, typically showing polyphase defor-
mation and structural repetition through faulting and folding. One of the
principal problems facing structural, stratigraphic, and tectonic synthesis
of greenstone belts lies in distinguishing between structural and strati-
graphic contacts in areas of poor exposure and in the near-absence of
unambiguous tools for relative age determination and correlation. Whereas
it was once fashionable to regard thick, apparently intact, uniformly
facing successions of volcanic and sedimentary rocks in greenstone belts as
forming coherent stratigraphic sections, often in excess of 15 km in
thickness, the present tendency is often to infer that such sequences, at
least on this planet, are composite, formed by the tectonic repetition of
considerably thinner stratigraphic sections.

The problem, now as previously, is the field recognition of faults,
particularly stratiform faults, such as thrusts. In the Barberton belt,
for instance, there are large areas, particularly in upper parts of the
succession, within which nearly stratiform thrust faults are present and
can be easily recognized using conventional means: (1) truncated and
offset stratigraphic units and folds, (2) unambiguously repeated strati-
graphic sequences, (3) the development of mylonitic and brecciated zones
along fault planes, and (4) the formation of drag folds in units adjacent
to the faults. However, throughout most of the classic sections of the
Onverwacht Group in the southern part of the belt, major faults identifiable
by such conventional criteria are absent. Although it has been suggested
that most of the apparent 12-km thickness of the Komati, Hooggenoeg, and
Kromberg Formations is an artifact of isoclinal folding of a much thinner
sequence (2), studies of facing directions throughout the section do not
bear out this interpretation (3). Arguments have also been advanced (2,
DeWit, this meeting) that chrome-mica-bearing alteration zones at the tops
of komatiitic units within this sequence represent stratiform shear zones
with displacements of perhaps 1-10 km. Unfortunately, however, these units
display none of the usual characteristics of faults (such as cross-cutting
relationships) and are developed only at the tops of komatiitic flows
(never at the tops of tholeiitic of felsic units). They exhibit cataclasis
and schistosity only where cross-cut by clearly later, through-going faults
or where present in areas where all units show penetrative deformation.
In most sections, these rocks display well-preserved, unsheared primary
spinifex and cumulate textures. Inferences that these zones represent
faults must at some point be based on a systematic consideration of their
characteristics, including clear enumeration of features indicating an

origin through faulting and the means of determining displacement.

Although it is clear that our ability to unambiguously differentiate structural and stratigraphic contacts in greenstone belts without fossils or rather fortuitous combinations of features will remain limited, the use of conventional criteria cannot be abandoned entirely. The possibility that thick, stratigraphically intact sequences are present in greenstone belts must remain as a working hypothesis until internal faults or folds can be identified based on clearly defined and well-understood criteria.

As noted above, it is our assessment that much of the controversy surrounding greenstone belt tectonics and evolution originates not from ambiguities in the genesis of rocks and structures in greenstone belts but from ambiguities in what those rocks and structures are and were. Future resolution of these controversies will rest more on careful, systematic studies of individual aspects of greenstone belts than on broad-brush syntheses or non-systematic collections of observations. A clear example of the success of the systematic approach is the role detailed geochronological studies have played in resolving the evolution of the late Archean Canadian belts. These studies (e.g. 5) have confirmed the existence of large-scale volcanic cycles within the Canadian greenstone belts and the existence of stratigraphic sections up to 10 km thick.

The results of any attempted overview of the similarities and differences among Archean greenstone belts depend significantly on how the term "greenstone belt" is defined. Presently used definitions (8) range from exceedingly broad (supracrustal successions in which mafic volcanic rocks are predominant) to relatively narrow (those requiring specific components, such as ultramafic or komatiitic lavas, and the increasingly common, largely implicit definition equating greenstone belts and ophiolites). Based on consideration of features common to most of the greenstone belts discussed in the present set of abstracts, we offer the following definition:

> Greenstone belt – an orogen made up largely of mafic to ultra-mafic volcanic rocks and their pyroclastic equivalents and epiclastic derivatives, showing intense macroscale deformation but regionally low grades of thermal alteration, and extensively intruded by penecontemporaneous or slightly younger granitoid plutons.

Virtually all terranes commonly considered as greenstone belts are encompassed by this definition, including many Phanerozoic examples. A critical aspect of this definition, and one that requires careful consideration, is that the terms "greenstone belt" and "ophiolite" are not synonymous. Rather, as in Phanerozoic orogens, ophiolites or ophiolite-like sequences may be components of greenstone belts.

Even with the restrictions imposed by this or most other definitions, greenstone belts constitute a highly diverse family of terranes. Some include an essentially continuous spectrum of komatiitic, tholeiitic, and calc-alkaline lavas, such as many belts in the Superior Province; others show a strongly bimodal volcanic suite (Barberton). Some are dominated by eruptive rocks (Superior Province, eastern Pilbara Block, and Barberton), others by sedimentary units (Slave Province and many Indian belts). The volcanic sequences in older greenstone belts (Barberton and eastern Pilbara) accumulated under shallow-water, anorogenic platform conditions; those in

most younger belts represent deep-water, tectonically active settings (7).
Additional differences have been noted by other investigators (9, 10).
These differences encompass nearly as much variability as represented by
the spectrum of modern orogens. A possible implication of this diversity
is that greenstone belts may represent tectonic settings as varied as those
represented by modern orogenic belts.

The results of most modern studies of greenstone belts suggest that
close scrutiny of individual belts usually allows identification of litho-
logically and structurally analogous modern terranes and, by inference,
tectonic settings. There is an emerging consensus, for instance, that the
petrologic, structural, and geochronological characteristics of large parts
of the Superior Province indicate that it is an assembly of late Archean
volcanic arcs formed along convergent plate boundaries that were basically
similar to volcanic arcs and convergent boundaries today (Card, this
volume). An important dissenting view, however, is expressed by David and
others (this volume). Parts or all of the volcanic sequences of other
Archean belts have been interpreted to represent oceanic or simatic crust
formed at spreading centers.

Using a similar argument, the more-or-less regular vertical strati-
graphic succession in greenstone belts, including lower volcanic and upper
sedimentary stages, is grossly similar to the stratigraphic sequences in
many modern orogens. If a genetic similarity is indicated, then it may be
expected that individual greenstone belts include rocks formed in an
evolutionary spectrum of tectonic settings. Perhaps, under ideal conditions
of preservation, these may range from cratonic rift and/or ocean floor
settings near the base to volcanic arc and, in some instances, cratonic or
peri-cratonic settings at the top.

At the same time, if we look closely at individual greenstone belts,
many features can be identified that are not present in their younger
analogs. These include the common presence of extensive komatiitic lavas,
banded iron formation, ocean-crust-like sequences (ophiolites) in excess of
10 km thick, and regionally extensive shallow-water sedimentary units
deposited in anorogenic simatic settings. Some of these features, such as
banded iron formation, reflect differences in modern and Archean systems
that are probably unrelated to tectonics. Others, such as unusually thick
ocean-crust sequences and widespread shallow water simatic platforms, may
reflect important differences between Archean and Phanerozoic tectonic
systems, if not in fundamental character then in local expression.

Future resolution of many of the outstanding controversies of green-
stone belt evolution rests in detailed systematic studies of (i) individual
properties of individual greenstone belts (structural style, alteration,
sedimentology, petrology), (ii) differences among Archean greenstone belts,
and (iii) similarities and differences between Archean belts and younger,
apparently analogous terranes.

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SEDIMENTOLOGICAL AND STRATIGRAPHIC EVOLUTION OF THE SOUTHERN PART OF THE BARBERTON GREENSTONE BELT: A CASE OF CHANGING PROVENANCE AND STABILITY;

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The sedimentological and stratigraphic evolution of the 3.5 to 3.3 Ga Barberton Greenstone Belt can be divided into three principal stages: (1) the volcanic platform stage during which at least 8 km of mafic and ultramafic volcanic rocks, minor felsic volcanic units, and thin sedimentary layers (Onverwacht Group) accumulated under generally anorogenic conditions, (2) a transitional stage of developing instability during which widespread dacitic volcanism and associated pyroclastic and volcaniclastic sedimentation was punctuated by the deposition of terrigenous debris derived by uplift and shallow erosion of the belt itself (Fig Tree Group), (3) an orogenic stage involving cessation of active volcanism, extensive thrust faulting, and widespread deposition of clastic sediments representing deep erosion of the greenstone belt sequence as well as sources outside of the belt (Moodies Group).

I. The platform stage of Barberton Greenstone Belt development is represented by rocks of the predominantly volcanic Onverwacht Group. Sediments deposited during this stage included (a) dacitic breccias, conglomerate, and coarse sands deposited as part of and adjacent to felsic volcanic centers and, less abundantly, proximal mafic lapillistones and tuffs; (b) distal felsic volcaniclastic and pyroclastic layers consisting mainly of fine ash, dust, and accretionary lapilli, (c) biogenic deposits such as carbonaceous oozes, carbonaceous muds, bacterial mats, and locally, stromatolites, and (d) orthochemical sediments including evaporites, barite, carbonate, and possibly siliceous deposits. The bulk of these sedimentary units show clear evidence of having been deposited under shallow-water conditions. The regional stratigraphic continuity and sedimentological integrity of sedimentary layers within this sequence, the predominantly shallow-water depositional setting, and the paucity of debris derived from the uplift and erosion of older rock sequences indicate that the overall depositional and tectonic setting was a broad, low-relief, shallow-water anorogenic platform (1).

II. Rocks traditionally assigned to the Fig Tree Group were deposited during a transitional phase of greenstone belt evolution. These are exposed in a complex succession of thrust sheets that provide numerous exposures of each part of the stratigraphic sequence (2). The lowest part of the Fig Tree is characterized by distal volcaniclastic units and carbonaceous cherts resembling those in the Onverwacht but showing rapid lateral facies changes. In particular, 40 to 50 m of predominantly carbonaceous chert in some structural belts can be correlated with a sequence of interbedded ultramafic lavas, banded cherts, carbonaceous cherts, stromatolites, and volcaniclastic units at least 500 m thick in other areas (2).

The overlying 200 to 500 m of rocks includes two principal components. By far the greatest thicknesses of Fig Tree strata consist of heavily altered dacitic pyroclastic and volcaniclastic detritus (3). This succession includes three main lithofacies: (a) plagioclase-phyric intrusive rocks that may locally grade into extrusive flows, (b) proximal, plagioclase-phyric breccias and conglomerates, probably developed as lava domes and surrounding coarse epiclastic units, and (c) regionally extensive ash

deposits, tuffs, and their current-worked equivalents, volcaniclastic sandstone and siltstone. The bulk of the finely laminated cherty ferriginous sediments characterizing Fig Tree rocks throughout much of the Mountain Land represent altered fine-grained dacitic volcaniclastic deposits. In contrast to previous interpretations, we consider the Fig Tree to represent a predominantly volcanic interval, perhaps more closely related petrogenetically to the Onverwacht Group than to the suprajacent orogenic Moodies succession.

Interbedded with these volcanic and volcaniclastic strata are thin, lenticular units of chert-pebble conglomerate and chert-grit sandstone showing rapid lateral facies changes and apparently representing debris derived from local uplifts within the greenstone belt. Most of the debris can be identified with underlying silicified rocks of the Fig Tree Group; there is little evidence for major uplift or deep erosion of the greenstone belt at this time.

III. Rocks which have traditionally been included within the Moodies Group represent three main clastic lithofacies: (a) a sequence of quartz-poor, highly altered sands and fine gravels derived by erosion of the subjacent dacitic rocks; (b) thick, coarse, chert-clast conglomerate and chert-grit sandstone derived by weathering and erosion of uplifted parts of the greenstone belt, and (c) quartzose and locally K-spar-rich sandstone representing the erosion of sources outside of the greenstone belt, possibly but not necessarily including the intrusive granitoid rocks and/or the Ancient Gneiss Complex or its equivalents.

Although the stratigraphic sections in most structural belts can be correlated with one another, there is as yet no satisfactory reconstruction of their original relative depositional positions. So-called northern facies rocks in the Mountain Land also belong to allochthonous terranes and their present location relative to units to the south is clearly of tectonic rather than depositional origin.

The overall sequence includes numerous minor unconformities and at least one major break. Within the Onverwacht Group, pauses in effusive activity are marked locally by weathering and erosion of flow surfaces, but no significant formation or accumulation of clastic debris. The inception of felsic volcanism both in the upper Hooggenoeg formation and the Fig Tree Group was accompanied by minor instability and local erosion of underlying rocks. Also, the formation of large, high-relief subaerial felsic volcanic edifices in Hooggenoeg and Fig Tree times was followed by extensive erosion and truncation of these complexes. The major structural unconformity within the Barberton sequence occurs locally at the base of the Moodies Group. Although a number of apparently conformable Fig Tree-Moodies transitions occur, over wide areas, the Moodies was deposited with angular unconformity on rocks as old as the Hooggenoeg Formation. This contact has additionally been complicated by structural movement.

The sedimentological development of the Barberton Greenstone Belt reflects three principal tectonic stages involving three contrasting sources of clastic sediment. The volcanic platform stage, represented by rocks of the Onverwacht and Fig Tree Groups, was primarily an interval of rapid effusion of lavas, subsidence, but little differential tectonic movement. The main sources of clastic detritus were first cycle, active, high-relief, felsic and, to a lesser extent, mafic volcanic centers. The second stage, represented by rocks of the Fig Tree Group, was one characterized by continuing, regionally extensive volcanism and developing

tectonic instability reflected by the presence of extensive lateral facies changes and small intra-platform uplifts that supplied shallow-level intraformational debris to local sedimentary systems. Latest Fig Tree and Moodies deposition was influenced by concurrent thrusting and orogenesis. Sediments were derived initially from both shallow and deep levels within the greenstone belt and, later, from distant quartz and K-spar rich sources outside of the belt.

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EVIDENCE FOR STRUCTURAL STACKING AND REPETITION IN THE GREENSTONES OF THE KALGOORLIE DISTRICT, WESTERN AUSTRALIA

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INTRODUCTION

Most previous stratigraphic interpretations of the southern part of the Norseman-Wiluna Greenstone Belt have proposed polycyclic sequences (e.g. 1-5). These invoked two and sometimes three successive suites of mafic and/or ultramafic volcanics and intrusives separated by felsic volcanics and immature clastic sediments, however no distinctive lithological differences were reported between successive mafic-ultramafic sequences. When interpretations (6, 7), further to the north, are integrated, a total of four separate major mafic-ultramafic suites emerges for a large part of the Norseman-Wiluna Belt. Although the author does not intend to imply that all polycyclic stratigraphies are wrong in principle such a situation seems suspiciously over-complex and stimulates the need to look critically at the individual areas where stratigraphies have been erected. For the Kalgoorlie area in the south, some of the schemes have already provoked scepticism (8, 9) and a simpler model consisting of one cycle subject to structural repetition has been evolved by workers in the Geological Survey of Western Australia (10) for part of this area. The latter authors drew attention to the 'carbon copy' similarity between the elements of some polycyclic stratigraphies. Much more regionally extensive integrated structural and stratigraphic data is still required to evaluate the relationship between structure and stratigraphy more fully, an objective substantially limited by poor outcrop and deep weathering, but with due effort, far from unattainable.

OUTLINE OF STUDY

Regional mapping by the author in an area of approximately 20,000 km^2 centred on Kalgoorlie revealed many problems and anomalies in several of the published stratigraphic schemes. However since insufficient critical stratigraphic and structural evidence had been given in support of the schemes it has not been easy to check the bases on which they were erected. The following lines of investigation have been pursued.

* Regional distribution and interrelationships of lithologically similar sequences previously regarded as distinct, based on mapping, mineral exploration data, and geophysical interpretation. Emphasis has been on the mafic-ultramafic suites because they are the most easy to define and map.

* Critical evaluation of contacts and their associated structural features.

RESULTS

a. **General features of the repetition** There are several instances where mafic-ultramafic suites previously proposed as younger (e.g. Coolgardie-Kurrawang area in 4) (see fig. 1) join or merge with their 'older' counterparts when mapped over various distances. They range in size from splinter-like splays a few kilometres long diverging from a major mafic belt by up to a kilometre, to extensive sheets which are traceable for tens of kilometres as separate entities before joining with and becoming indistinguishable from their 'older' counterparts. Some successions are isolated in metasedimentary terrain, and never connect with their sequences of origin; however this situation is unusual. In areas where like elements of two proposed cycles are juxtaposed or interconnected (e.g. Widgiemooltha and Spargoville areas in map of 5) there seems to be no clear reason why they should have been regarded as separate.

EVIDENCE FOR STRUCTURAL STACKING AND REPETITION
J. E. Martyn

The apparent stratigraphic thicknesses of many of the previously proposed younger mafic-ultramafic sequences is very variable. While they may be measurered in kilometres in some areas, in many localities the sequences are attenuated and deformed. They may be traced for tens of kilometres as apparently conformable packages of all or most of the major mafic and ultramafic lithologies, though individually these lithologies may occur as lenses or sheets hundreds or even only tens of metres in thickness. While such sequences have been interpreted in the past as volcanic intercalations in a eugeosynclinal sedimentary pile (2, 4), or the beginnings of new volcanic cycles, their degree of deformation and tendency to be smaller scale carbon copies of their 'older' counterparts is more consistent with structural repetition. In a number of instances they are overlain by felsic volcanic rocks suggesting cyclic development in a uniformly facing sequence. This is here regarded as evidence that repetition has been mainly by faulting and not by recumbent or isoclinal folding.

b. Observations on Contacts

Many previous stratigraphies (e.g. 3-5) have been erected in areas where fragmentary facing evidence suggests thick uniformly facing sequences. The potential for strike dislocations has generally been overlooked despite heterogeneous shear deformation. There has been an absence of critical treatment of major formational contacts to establish whether they are normal or tectonised. This is understandable in some instances since such contacts are rarely well exposed, however diligent search by the author has revealed many key outcrops. The vast majority of these provide compelling evidence that all is not well with the published polycyclic stratigraphies. Examination of contacts, especially basal ones, and the contact areas of the previously proposed younger mafic-ultramafic suites, commonly reveals strong peneconcordant shearing, recrystallised mylonitic or other cataclastic rocks, or in one instance (the Kalpini formation which is the highest mafic-ultramafic suite of the interpretation in 3), an overturned but undeformed contact with clear facing evidence the reverse of that previously proposed. Mapping the relationship between the contact zones and primary layering often reveals subtle discordances not readily explained by unconformity.

MODEL

Many proposed structural repetitions or fault slices are linear, others are arcuate and folded around major upright structures. Linear belts are often controlled by throughgoing transcurrent deformation zones with pronounced sub-horizontal lineations. Arcuate systems however were conceivably generated by earlier processes such as thrusting or gravitational gliding predating upright folding. Although transcurrent shearing is a feasible mechanism for repetition for at least some of the more linear belts, it is possible that even many of these began life as early thrust sheets and became stretched and aligned by later transcurrent deformation. Early thrusts, recumbent folds and layer-parallel shear fabrics have been documented in several localities in the Norseman-Wiluna Belt where prevailing strikes deviate from the NNW regional grain, or where tight upright folding is subdued or absent (e.g. 11-16). In one instance (11) a narrow mafic-ultramafic belt in sediments has clearly been generated by an overthrust. Almost certainly the recognition of this structural style in east-west trending or gently domed areas is a consequence of preservation. It is undoubtedly present also in NNW trending linear domains but is overprinted and hard to recognise. It is emphasised that thrust repetition does not explain all of the previously proposed younger cycles in the district. Some are a consequence of misinterpretation, by placing too great a significance on isolated stratigraphic facing observations, or from attempts to correlate across major upright faults. Broad regional observations by the author suggest that thrust repetition may be much more strongly developed in the Kalgoorlie district than elsewhere in the Norseman-Wiluna Belt though this conclusion is tentative.

Thrusting does not appear to have occurred on a scale comparable with many Phanerozoic convergent plate boundaries. There is no evidence of juxtaposition of strongly contrasting domains, or of high pressure metamorphism. There is also a lack of pronounced east-west assymetry across the Norseman-Wiluna Belt as a whole. The tectonics can be viewed more in terms of a rearrangement of familiar elements of the local stratigraphy, a situation more consistent with a closed or intracratonic setting, rather than an open plate margin. This accords with models such as those of 17. As such, intrabasinal gravity gliding resulting from early uplift heralding later vertical tectonic events is the most favoured model by the author. This is consistent with the sedimentation style which is dominated by turbidites and includes debri flow deposits. Olistostromes have also been reported (Taylor, in 18). In some respects the scheme resembles one proposed for the Barberton Greenstone Belt (19). Felsic volcanism was intimately associated with sedimentation, and it is possible that concomitant granitic intrusion into a dense sheet of mafic-ultramafic volcanics may have triggered the instability that first led to the sedimentation and later to gravity gliding tectonics. Subsequent folding and faulting of the tectonically stacked sequence would have created the illusion of a polycyclic sequence which has suffered only upright folding and shearing. The upright tectonic events have generated their own set of interpretive problems. Peneplenation, and Tertiary lateritic weathering ultimately obscured much of the important evidence. The proposed tectonic history is summarised in fig. 2.

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FIG. 1

Fig. 1
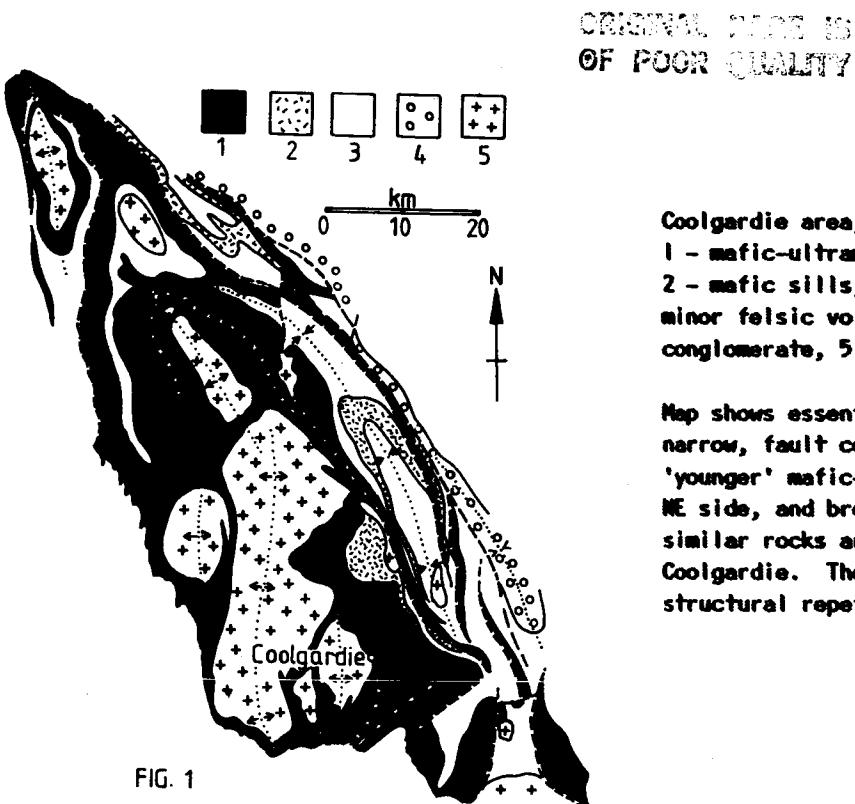
Coolgardie area, 40km SW of Kalgoorlie:
1 - mafic-ultramafic volcanics,
2 - mafic sills, 3 - sediments and
minor felsic volcanics, 4 - polymict
conglomerate, 5 - granitoid intrusives.

Map shows essential continuity between
narrow, fault controlled strips of
'younger' mafic-ultramafic rocks along
NE side, and broader areas of 'older'
similar rocks around and to NW of
Coolgardie. The former are clearly
structural repetitions of the latter.

FIG. 2
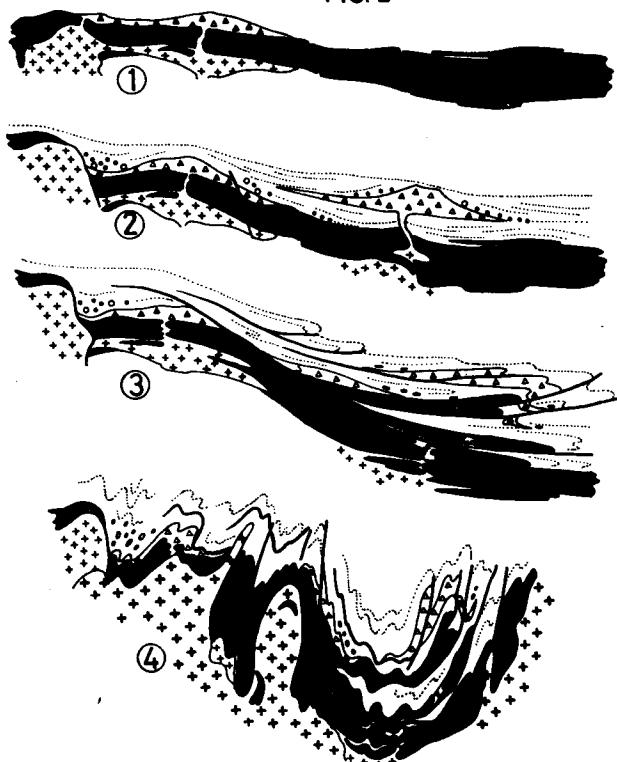
Fig. 2

Schematic representation of sequence of
events producing structural repetition.

1. Extensive komatiitic to tholeiitic
volcanism is followed by localised
felsic eruption and related intrusion.

2. Uplift and subsidence leads to
rapid deposition of immature
volcanogenic sediments. This is
accompanied by further felsic volcanism
and intrusion and destabilisation of
basin.

3. Further uplift induces gravity
gliding and thrusting leading to
structural stacking.

4. Upright folding and faulting and
batholithic intrusion produces complex
terrain. Effects of older thrusting
may be mistaken for polycyclic
stratigraphy. Upright shears induce
additional complexity.

PRELIMINARY STRUCTURAL MODEL FOR THE SOUTHWESTERN PART OF
THE MICHIPICOTEN GREENSTONE BELT, ONTARIO; George E. McGill and
Catherine H. Shrady, Dept. of Geology and Geography, University of
Massachusetts, Amherst, MA 01003

The southwestern part of the Michipicoten Greenstone Belt includes
a 100 km^2 fume kill extending northeastwards from the town of Wawa,
Ontario. Except for a strip along the Magpie River that is covered by
Pleistocene gravels, outcrop in the fume kill averages about 30-50%.
Within this area are all the major lithologic belts characteristic of
the southwestern fourth of the Michipicoten Greenstone Belt. All of
the area mapped to date lies within Chabanel Township, recently mapped
at 4" = 1 mile by Sage et al. (1). Following a brief reconnaissance
in 1983, mapping at a scale of 1" = 400' was begun within and adjacent
to the fume kill in 1984. We have concentrated on two objectives:
1) determination of the geometry and sequence of folding, faulting,
cleavage development, and intrusion; and 2) defining and tracing litho-
logic "packages", and evaluating the nature of the contacts between
these packages. Results for objective 1) are discussed in a companion
abstract (2); this abstract will present tentative results for objective
2).

The entire Michipicoten Greenstone Belt has experienced relatively
late movement on steep faults, most of which trend approximately NNW
or NE (1,2,3). Some of this movement preceded the emplacement of diabase
dikes, some followed. These displacements may be easily removed in
order to reassemble older structures, which are of much greater tectonic
interest.

For mapping and descriptive purposes, it long has been customary
to divide the stratified rocks of the Michipicoten Greenstone Belt
into 4 major lithologic groups (1,3): mafic-intermediate volcanics,
intermediate-felsic volcanics, clastic sediments, and chemical sediments
(including iron formation). This is certainly valid, because outcrop
belts of these groups maintain integrity for long distances. However,
there are along-strike intergradations among them, and there is no
easy way to correlate between physically separated belts of similar
lithology. This last problem means that there is no really dependable
belt-wide stratigraphy, and relative ages of the various belts of similar
lithology are known only in the few places where modern radiometric
ages have been measured (4,5).

Our detailed mapping (Fig. 1) indicates that the situation is
more complex than one would infer from published maps and descriptions
(1,3,6). There are several lithologic packages within the single belt
of clastic sediments in Chabanel Township, all of which appear to be
bounded by fault contacts. In some cases, stratigraphic way up reverses
across these faults, in other cases it does not. At map scale, the
package boundaries follow bedding or volcanic layering on one or both
sides, but locally this is not so, and at outcrop scale it commonly

STRUCTURAL MODEL, MICHIPICOTEN BELT
McGill, G.E. and Shrady, C.H.
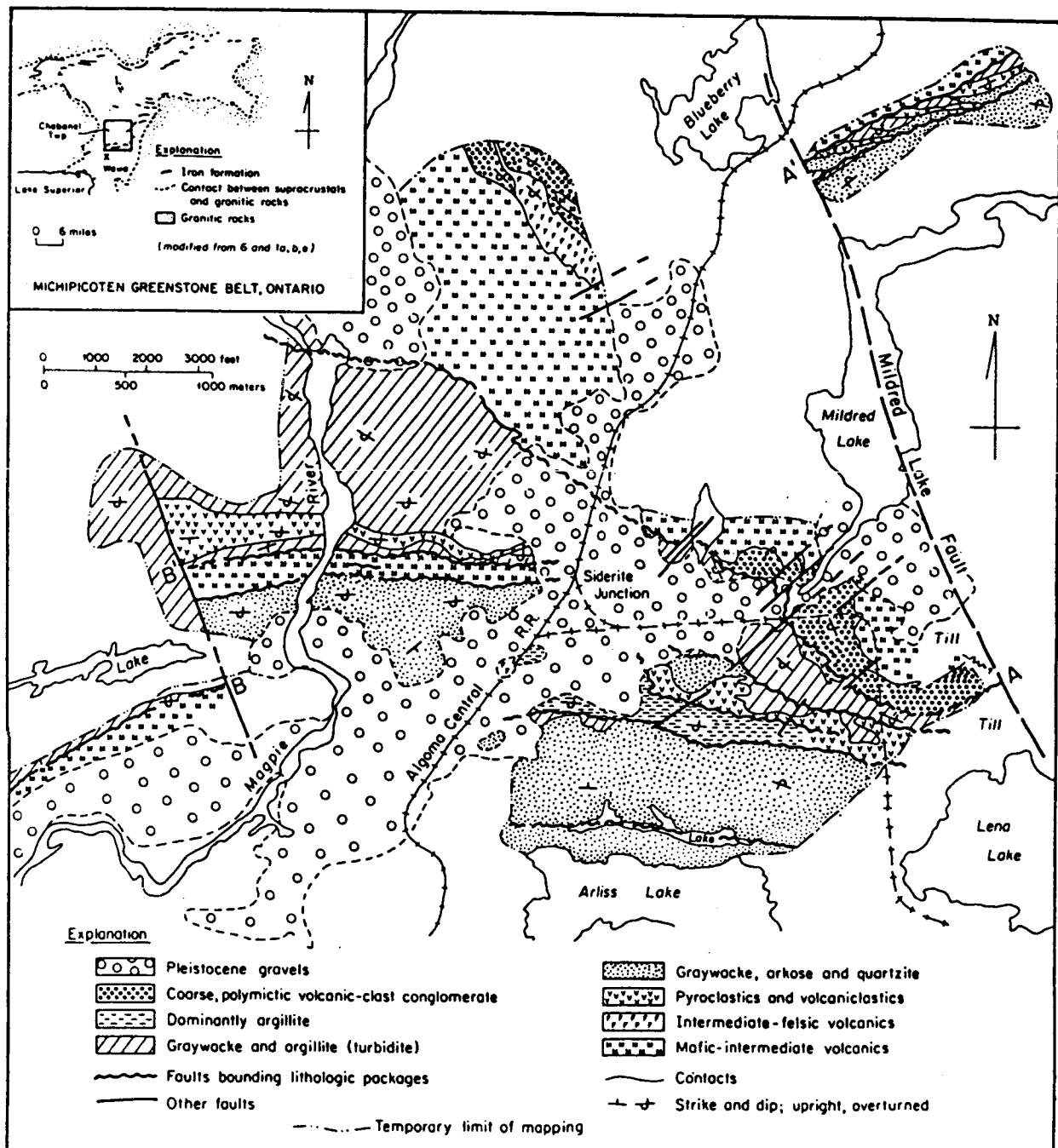
Fig. 1. Geologic sketch map of the central part of Chabanel Township,
Ontario. All intrusive igneous rocks omitted for simplicity.
B-B' and A-A' indicate corresponding points across late faults.

STRUCTURAL MODEL, MICHIPICOTEN BELT

McGill, G.E. and Shrady, C.H.

is not so. In places, these faulted boundaries are characterized by locally developed cleavages, excessive flattening or elongation of pebbles, or minor folds.

The area we have mapped seems to be a zone of faults and folds separating a large region to the south underlain by overturned rocks with tops north from an even larger region to the north underlain by overturned rocks with tops south (1,6). This relationship would seem to indicate an antiformal fold in the inverted limb of a very large nappe, but we have not been able to define such a structure, and rocks that should correlate across the structure are not the same age (R. Sage, pers. com.). Major faulting thus is necessary, but earlier or synchronous folding at township or larger scale would seem necessary to account for the opposed overturning. Almost all of the rocks north and south of our area are volcanic, so it may never be possible to determine if these terranes consist of continuous sections or if they, too, are divided into fault-bounded packages.

Because we have yet to sort out the sequence of minor and major structures with sufficient confidence, and because completed detailed mapping covers such a small fraction of the total belt, we prefer to be rather conservative about interpreting our data. Key observations include a "stratigraphy" that consists mostly of fault-bounded "packages", the apparent early age of these faults, and the large areal extent of the inverted sequences facing each other. The most attractive and probably the simplest explanation for these relationships involves early imbricate thrusting--before the imposition of the almost universal steep dips. However, this interpretation remains to be proved.

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THERMAL IMPLICATIONS OF METAMORPHISM IN GREENSTONE BELTS AND THE HOT ASTHENOSPHERE-THICK CONTINENTAL LITHOSPHERE PARADOX; Paul Morgan, Department Geosciences, Purdue University, West Lafayette, IN 47907.

From considerations of secular cooling of the Earth and the slow decay of radiogenic heat sources in the Earth with time, the conclusion that global heat loss must have been higher in the Archean than at present seems inescapable. The mechanism by which this additional heat was lost and the implications of higher heat loss for crustal temperatures are fundamental unknowns in our current understanding of Archean tectonics and geological processes. Higher heat loss implies that the average global geothermal gradient was higher in the Archean than at present, and the restriction of ultramafic komatiites to the Archean and other considerations suggests that the average temperature of the mantle was several hundred degrees hotter during the Archean than today (1). In contrast, there is little petrologic evidence that the conditions of metamorphism or crustal thickness (including maximum crustal thickness under mountains) were different in Archean continental crust from the Phanerozoic record (see 1). Additionally, Archean ages have recently been determined for inclusions in diamonds from Cretaceous kimberlites in South Africa (2), indicating temperatures of 900 to 1300 degC at depths of 150 to 215 km (45 to 65 kbar) in the Archean mantle (3), again implying relatively low geothermal gradients at least locally in the Archean. In this contribution the thermal implications of metamorphism are examined, with special reference to greenstone belts, and a new thermal model of the continental lithosphere is suggested which is consistent with thick continental lithosphere and high asthenosphere temperatures in the Archean.

High-grade metamorphism is common in Archean terrains (4, 5), and includes some greenstone belts, such as in the Yilgarn block of SW Australia (6). High metamorphic temperatures (700 degC or more) and often high metamorphic pressures (5 to 10 kbar or greater) are indicated by the mineral assemblages in these terranes, and they are underlain in most cases by continental crust of normal thickness (7, 8). Conductive thermal relaxation models have been proposed to predict the thermal conditions of metamorphism in the crust following tectonic activity such as underthrusting (e.g., 9-11). As demonstrated by Ashwal and Morgan (7), however, simple thermal relaxation of thickened crust cannot reasonably produce the high temperatures required by granulite metamorphism with a thick section of crust (30 km or more) below the shallowest depth of granulite metamorphism without requiring the lower part of the crust to be supersolidus. Basically the temperature range for granulite metamorphism is so close to estimates of the crustal solidus for reasonable crustal compositions (e.g., 12), that a positive geothermal gradient below the shallowest depth of granulite metamorphism causes the geotherm to intersect the solidus above the Moho. Ashwal and Morgan (7) conclude that unless granulite metamorphism occurs only near the base of the crust and the thick section of crust now below the exposed granulites was added after metamorphism, major crustal magmatic activity is associated with granulite metamorphism. Such extreme thermal conditions are not required by lower grades of metamorphism, but any metamorphic gradients which indicate a high geotherm suggest the upward transport of heat by magma unless the crust is thin.

If it is accepted that magmatic heat transport is an essential component of the crustal thermal regime during the peak thermal conditions recorded by the metamorphic mineral assemblages in the crust (at least where high geothermal gradients are indicated), then maximum temperatures recorded in

these systems were buffered by the solidus. The occurrence of young granulites at the top of sections of normal thickness crustal sections similarly indicates that modern maximum geothermal gradients are buffered by the solidus. A similar conclusion is indicated by heat flow data from areas of recent tectonism in which high heat flow must result from magmatic heating of the crust (e.g., 13). Maximum temperatures at shallow depth are buffered by the boiling point curve at hydrostatic or lithostatic pressures, below which maximum temperatures are buffered by the crustal solidus. As these maximum crustal temperatures are commonly encountered in areas of active tectonism and magmatism today, it is impossible for maximum temperatures recorded by Archean metamorphic assemblages to have been higher than modern maximum temperature conditions unless the solidus was different. Thus, in this buffered system, higher heat loss in the Archean is not expected to be recorded by metamorphic assemblages indicating higher geothermal gradients than peak modern conditions, although these peak crustal thermal conditions may have been more widespread in the Archean than at present.

The occurrence of high-grade (granulite) metamorphism in Archean greenstone belts suggests that either the high-grade areas were produced near the base of the crust and subsequently the crust has been thickened below the high-grade terranes, and/or magmatism was an important process during the high-grade metamorphism. The intimate association of plutons with the greenstone belts in "granite-greenstone" terranes suggests the importance of magmatism during this high grade metamorphism, and is consistent with models which suggest basal melting of stacked simatic thrust sheets during the evolution of at least some greenstone belts (14-16).

Perhaps the most paradoxical indicator of Archean thermal conditions with respect to higher global heat loss is the relatively low Archean geothermal gradients indicated by the formation of diamonds of Archean age. The diamond stability field is consistent with geotherms predicted for modern shield areas with thick (150 km or greater) lithosphere (e.g., 13). Meyer (3) has suggested that diamonds were formed in the asthenosphere which in turn suggests that perhaps the higher temperatures deduced for the Archean mantle from the occurrence of komatiitic lavas were not universal. A more common interpretation of the diamond data is that they indicate the existence of thick "keels" of subcontinental lithosphere below at least some areas during the Archean (1, 16). However, as the lithosphere is intimately related to the thermal boundary of upper mantle convection, it would be expected that this boundary layer and the lithosphere would have been thinner during the Archean with higher global heat loss and mantle temperatures. A possible solution to this paradox may be found in the intrinsic heat production of continental lithosphere.

There are two basic variable parameters that control the stable thickness of the continental thermal boundary layer (lithosphere), the heat production within the layer and the heat input to its base (13, 17). The layer thins if heat input to its base increases, and thickens if the heat input decreases. This heat input depends upon the temperature difference between the lower portion of the stable boundary layer and the underlying convection cell, or more specifically the temperature gradient in the lowest portion of the layer. As this gradient decreases to zero, the heat input to the base of the lithosphere decreases to zero (negative gradients are not permissible in a stable thermal boundary layer). The thickness of stable continental lithosphere with zero heat input at its base is independent of the global heat loss, assuming that the heat can be lost elsewhere (oceanic and other continental lithosphere), and this may possibly be a mechanism for maintaining

thick continental lithosphere at a time of high global heat loss and high
average mantle temperatures.

The condition for zero heat flux into the base of the stable continental
lithosphere is that the temperature increase within the lithosphere due to its
intrinsic radiogenic heat production creates a geotherm that is asymptotic to
the asthenosphere isotherm (or adiabat with an adiabatic basal heat flux).
For thick lithosphere this condition requires a small but significant
component of heat production in the mantle lithosphere, and an example of such
a heat production distribution and geotherm are given in Figure 1. This
condition has the interesting property that thicker lithosphere is indicated
for higher asthenosphere temperatures for similar heat production
distributions. If heat production distributions of this type are realistic it
is unlikely that they are accidental (see also 18), and the concentration of
radiogenic heat production into the lithosphere by metasomatism and crustal
building processes may be related to the stabilization of continental
lithosphere.

Figure 1. Example of continental
lithosphere geotherm asymptotic
with asthenosphere isotherm as a
result of its intrinsic radiogenic
heat generation. A two component
crustal heat generation model is
assumed for this geotherm: An upper
crustal component decreasing
exponentially with depth from 2.7
μW/m**3 at the surface with a
depth scale length of 7 km, and an
additional uniform component of
0.09 μW/m**3 (geotherm model
modified from 19).
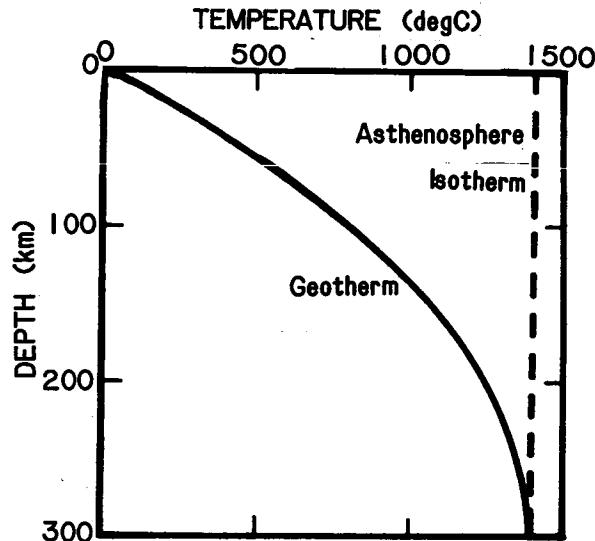
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623-634.

GEOCHEMICAL CHARACTERS AND TECTONIC EVOLUTION OF THE CHITRADURGA SCHIST BELT: AN ARCHAEAN SUTURE (?) OF THE DHARWAR CRATON, INDIA. S.M.NAQVI, NATIONAL GEOPHYSICAL RESEARCH INSTITUTE, HYDERABAD, INDIA.

The Chitradurga schist belt extending for about 450 km in a NS direction and 2-50 km across, is one of the most prominent Archaean (2.6 b.y.) tectonic features of the Indian Precambrian terrain, comprising about 2 to 10 km thick sequence of volcanosedimentary rocks. The basal unit of this belt is composed of an orthoquartzite-carbonate facies, unlike many other contemporary greenstone belts of the Gondwana land which begin with a basal mafic-ultramafic sequence. Eighty percent of the belt is made up of detrital and chemogenic sediments, their sucession commencing with a poorly preserved quartz pebble basal conglomerate and current bedded quartzites which, in turn, rest on tonalitic gneisses, the latter having been further remobilized alongwith the schist belt. Deposition of current bedded mature arenites indicate the existence of platformal conditions near the shore line. Polymictic graywacke conglomerates, greywackes, shales, phyllites, carbonates, BIFs (oxide, carbonate and sulfide) BMF's (Banded Maganese Formations) and cherts thus constitute the main sedimentary rocks of the belt. The polymictic conglomerates contain debris of rocks of older greenstone sequences, as well as an abundant measure of folded quartzites, BIF's and gneissic fragments which represent earlier orogenies.

Four different types of greywackes are recognised in the belt from N to S. Most of these have been derived from the surrounding tonalitic gneisses which contained older greenstone sequences as enclaves of various dimensions. However, the younger sequences in the north contain debris from the intrabasin volcanism also. The K-granites and gneisses are found to be progressively abundant in the source area of these graywackes as indicated by the granitic component of the debris of the younger graywackes sequences. Their REE patterns are characterized by both positive and negative Eu anomalies, the latter especially in the interbedded shales with grewackes. Geochemistry of the graywackes and chemogenic sediments thus indicate their deeper oceanic environment of formation. Although stratigraphic relation between the shallow water and deeper water sediments is uncertain, the basal orthoquartzites-carbonate sequences indicating platformal environment perhaps represent a facies change due to shallow water conditions along the shore line, and the greywacke suite those of deeper water away from it. Similar facies change is observed in the BIF's from shallower oxide to deeper sulfide facies.

The ultramafic rocks, mostly found in the lower sections of the belt, show pillow structures and spinifex texture and are komatiitic in composition. The mafic, intermediate and acid volcanics are found as detached outcrops in presumably higher

CHITRADURGA BELT
NAQVI S. M.

stratigraphic sections and show tholeiitic and calc alkaline affinities, probably produced by 5-15% melting. The ultramafic lavas were produced by deeper mantle melting source, the geochemical characteristics belonging to the oceanic class.

Most of the rock suites in the belt have been metamorphosed to greenschist facies. However, its eastern margin is found to be in thrust contact with the higher amphibolite facies rocks (700°C at 6-7 Kbr), and the southern part near Mysore consist of predominantly ultramafic rocks metamorphosed to amphibolite and granulite facies. The northern part of the belt near Gadag is least metamorphosed. Irrespective of the grade of metamorphism or of inferred ages of the various stratigraphic groups, the belt shows a remarkable structural homogeneity of 3 phases of deformation from N to S and E to W and a convexity towards East. Both major and minor Fl folds are tight isoclinal with shallow to steep plunges and subvertical to subhorizontal axial planes. The variation in the attitude and orientation of the Fl axes has been controlled by the F2 episode which has coaxially folded both the subparallel bedding and the first generation axial plane schistocity cleavage. Only at Fl hinges the intersection between S1 and S2 is discernible. F3 is found as general warps on F2 limbs. The Fl axial plane schistocity cleavage and F2 crenulation charge are generally dipping (horizontal to subvertical) towards the east. High grade rocks on the eastern margin have been thrust westwards over the low grade central part. Structural data indicate considerable crustal shortening along the belt. Inversion of stratigraphic sequence is reflected, at many places by the younging directions obtained from current bedding, graded bedding and pillow couvexities. Horizontal compression and collision tectonics therefore, appear to have played a significant role in the development of the structural configuration of the belt.

As the 3000 m.y. old gray banded gneisses, found on the eastern and western sides of the Chitradurga schist belt are similar, the existing observations suggest the following two possible models: i) The belt developed in a rift on the juvenile Archaean continental crust which collapsed upon loading by sediments, resulting in a shallow subduction and horizontal compression. (ii) The belt evolved on an "Oceanic" crust between two juvenile continental blocks to the East and West. Shallow subduction and horizontal movement of the Eastern block would then result in the present structural geometry and consequent welding of the two along this probable suture.

GEOCHEMICAL AND ISOTOPIC CONSTRAINTS ON THE TECTONIC SETTING OF THE SERRA DOS CARAJAS BELT, EASTERN PARA, BRAZIL; Wm. J. Olszewski, Jr., Dept. of Earth Sciences, Univ. of New Hampshire, Durham, NH 03824; Allan K. Gibbs and Karl R. Wirth Institute for the Study of the Continents, and Dept. Geol. Sci., Cornell University, Ithaca, NY 14853

The lower part of the Serra dos Carajas belt (Fig 1) is the metavolcanic and metasedimentary Grao Para Group (GPG) (1-6). The GPG is thought to unconformably overlie the older (but undated) Xingu Complex, composed of medium and high-grade gneisses and amphibolite and greenstone belts. The Lower Metavolcanic Sequence of the Grao Para Group (LMS) is estimated to be about 4-6 km thick, consisting of massive, vesicular, and porphyritic mafic volcanic flows and agglomeratic breccias and about 10-15% massive, flow-banded, brecciated, and tuffaceous prophyritic rhyolite (6). The LMS is overlain by the extensive, 100-400 m thick, and high-grade banded iron formations of the Carajas Formation, followed by an Upper Sequence (US) of 1-3 km of mixed volcanic and clastic and chemical sedimentary rocks. The stratigraphy of the US is poorly known, but it is thought to contain some quartz-rich arenites, suggesting mature continental provenance (6). Much thicker quartz-rich sandstones and conglomerates overlie the Upper Sequence, with unknown degree of conformity.

Petrographic, geochemical, and isotopic analyses of the bimodal metavolcanics of the LMS show these to be basalts, basaltic andesites, trachyandesites (shoshonitic), and rhyolites (6,8). Spilitic alteration is locally apparent, but the coherence of alkali element ratios and readily-altered trace element compositions suggests that most samples did not undergo strong alteration. Good correlation between HREES, Ti, and magnesium number in the mafic rocks demonstrate the effects of fractional crystallization in the mafic rocks. LREES, Si, K, Rb, Cs, and Ba do not correlate with magnesium number, suggesting that variable enrichments of these elements (fig. 2) reflect variable contamination of the basaltic melts with crustal material. Several contamination components must have been involved, since these elements are only weakly correlated among themselves, and with U, Th, Nb, and Ta. Rhyolite patterns show significant negative Eu anomalies.

Zircons from two quartz porphyritic rhyolites give an age of 2758 + 39 Ma (7), the best estimate of the age of eruption of the LMS. Rb-Sr whole-rock analyses of mafic rocks yield an isochron of 2687 + 54 Ma, similar within the range of calculated errors of the zircon age. Thus the GPG's Late Archean age is well established. The high initial Sr isotopic ratio 0.7057 for the mafic rock isochron is significantly higher than values of CHUR (0.7012, 9) or depleted mantle (0.7008, 10) for 2758 Ma. This indicates contamination by older continental crust. Sm-Nd results are too restricted in distribution to yield a usable isochron.

e Sr vs. e Nd values (Fig. 3) show a cluster around e Sr +50 and e Nd +3. These indicate that the magma was more likely derived from a depleted source than from a CHUR-like source. The high e Sr values are probably either the results of seawater interaction, leaving Nd isotope ratios intact; or contamination with older, presumably mafic crust that had elevated Rb/Sr ratios, but mantle-like Sm/Nd ratios. One rhyolite has similar e Nd and e Sr values, suggesting derivation from similar sources by similar processes. Three of the mafic samples have negative e Nd and positive e Sr values, possibly indicating contamination by older granulitic and granitoid crust. Note that the ranges of diversity in the e Sr and e Nd data can be seen in the basalts alone: the isotopic variation does not correlate directly with silica content. Diverse sources of contamination are indicated, and might be found

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in the diverse lithologies of the underlying Xingu Complex.

The geochemical data indicate that the GPG has many features in common with ancient and modern volcanic suites erupted through continental crust. The mafic rocks clearly differ from those of most Archean greenstone belts, and modern MORB, IAB, and hot-spot basalts. The geological, geochemical, and isotopic data are all consistent with depostion on continental crust, presumably in a marine basin formed by crustal extension. The isotopic data also suggest the existence of depleted mantle as a source for the parent magmas of the GPG. The overall results suggest a tectonic environment, igneous sources, and petrogenesis similar to many modern continental extensional basins, in contrast to most Archean greenstone belts. The Hammersley basin in Australia and the circum-Superior belts in Canada may be suitable Archean and Proterozoic analogues, respectively.

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SAMPLE LOCATIONS

SAVANNAS

GRANITE (1.8 Ga)

SANDSTONE UNIT

IRON FORMATION

GRAO PARA GROUP

XINGU COMPLEX

SERRA DOS CARAJAS, BRAZIL
Olszewski, W.J. Jr., Gibbs, A.K., and Wirth, K.R.
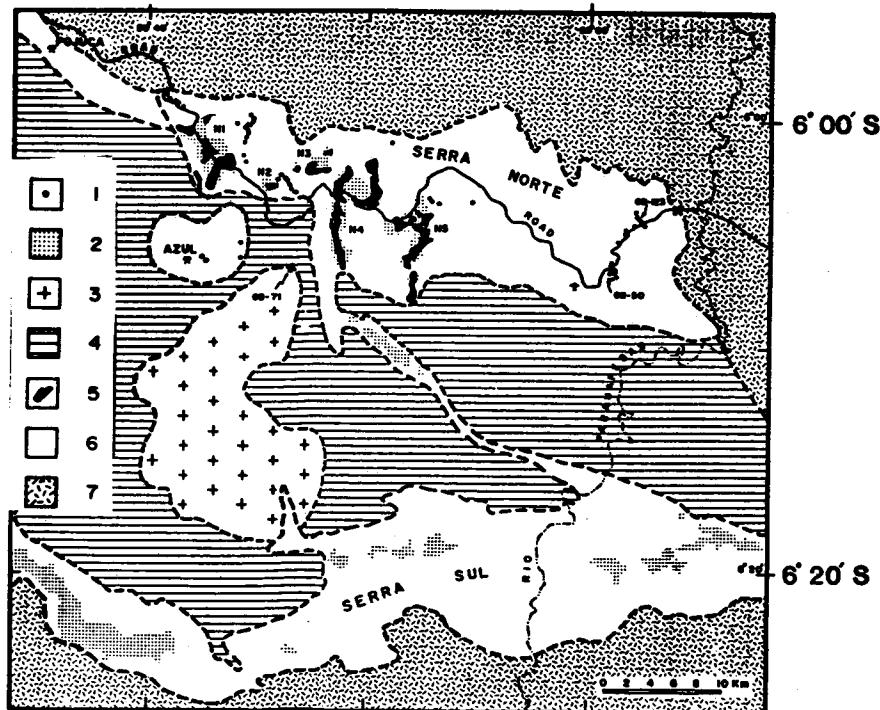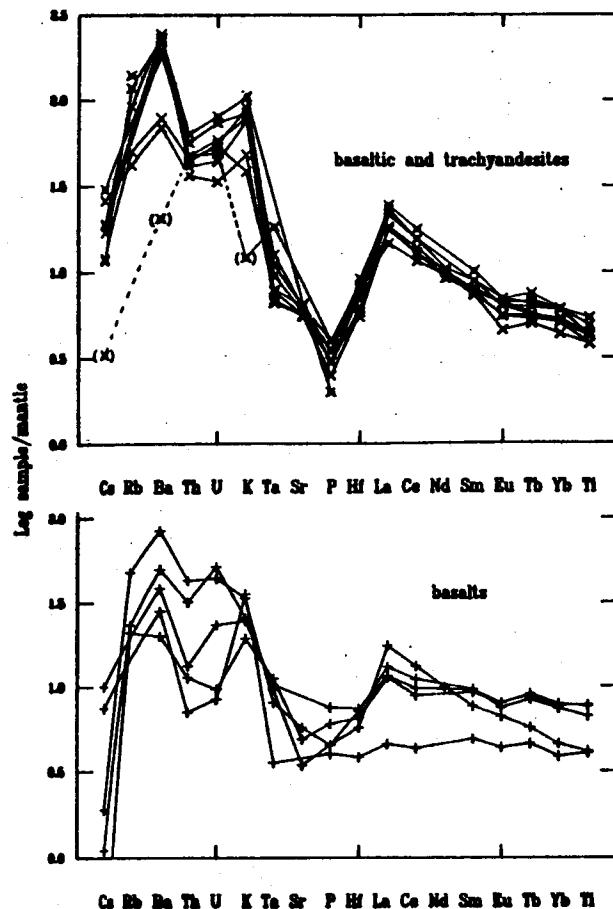
Fig. 2: Incompatable element diagrams for mafic rocks, normalized for model primitive mantle of Wood et al. 1979, and Hf, Eu, and Yb interpolated from chondrite data. Karoo basalts have similar enrichment patterns.

Fig. 3: Sr - Nd diagram for Grao Para Group metavolcanic rocks.
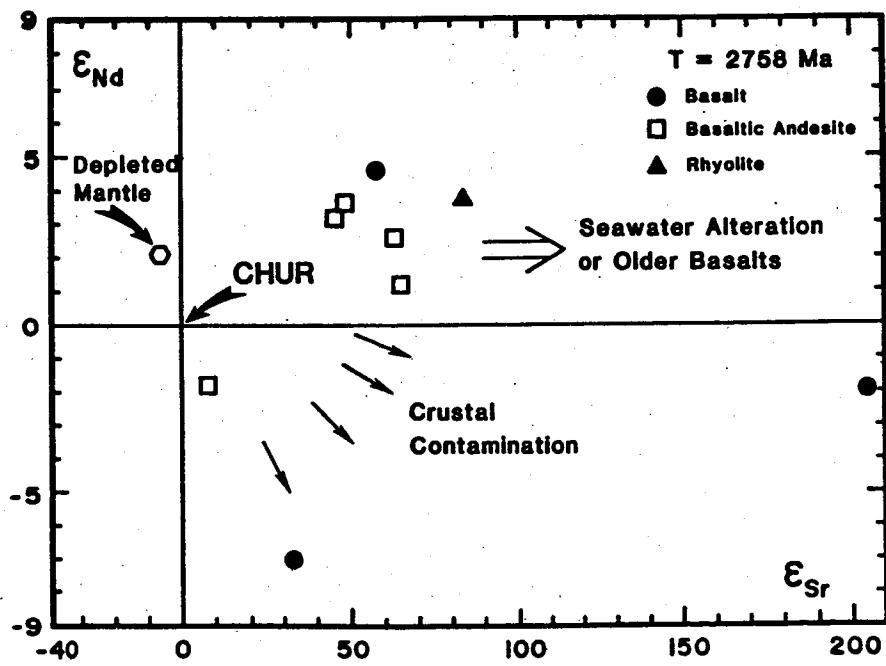
POLYPHASE THRUST TECTONICS IN THE BARBERTON GREENSTONE BELT.

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In the circa 3.5 by old Barberton greenstone belt, the supracrustal rocks form a thick and strongly deformed thrust complex. Structural studies in the southern part of the belt have shown that 2 separate phases of over-thrusting (D_1 and D_2) successively dismembered the original stratigraphy. Thrust nappes were subsequently refolded during later deformations (D_3 and D_4). This poster deals with the second thrusting event which, in the study region appears to be dominant, and (unlike the earlier thrusting), affects the entire supracrustal pile.

The supracrustal rocks form a predominantly NE/SW oriented, SE dipping tectonic fan (the D_2 fan) in which tectonic slices of ophiolitic-like rocks are interleaved with younger sedimentary sequences of the Diepgezet and Malalotcha Groups[1] (Fig. 1). Two distinct levels of decollement can be distinguished within this fan: (1) Within the ophiolitic sequence, usually below the pillow lavas. These zones are delineated by strongly sheared serpentinite lenses and talcose schists. Asbestos fiber is commonly developed in such sheared lenses, as for example in the Havelock and the Msauli asbestos deposits. (2) At the base of the Diepgezet Group, within ferruginous shales and banded cherts. This upper decollement zone is not always obviously sheared, but it is ubiquitously folded in a disharmonic manner and is thought to have been gravity induced, on a dynamic slope, during sedimentation, because: (1) The finely laminated rocks at this stratigraphic level are conformably to unconformably overlain by a 2 to 3 km thick medium to coarse grained clastic sequence (the rest of the Diepgezet Group and the Malalotcha Group; the Malalotcha Group is derived from a quartz-rich source and from the reworking of folded Diepgezet Material). (2) Within the D_2 fan, individual tectonic units may be folded independantly of one another (Fig. 2). The D_2 folds are mostly isoclinal, with fold axes broadly parallel to the thrust contacts (Fig. 2), and are contemporaneous with the emplacement of the nappes. Another set of D_2 folds is contemporaneous with the deposition of the Malalotcha Group sediments and probably formed in tectonically ponded basins, during periods of thrust propagation along the lower decollement level.

Structural and sedimentological data indicate that the D_2 tectonic fan was formed during a prolonged, multi-stage regional horizontal shortening event during which several types of internal deformation mechanisms were successively and/or simultaneously active. Movement appears to have been predominantly to the NW and to the N. During D_2, periods of quiescence and sedimentation followed periods of thrust propagation. Although the exact kimematics which led to the formation of this fan is not yet known, paleoenvironmental interpretations together with structural data suggest that D_2 was probably related to (an) Archean collision(s).

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THRUST TECTONICS IN THE BARBERTON BELT
Paris, I.

FIG 1

LEGEND.
- Malalotcha Group
- Diepgezet Group (proximal facies)
- Diepgezet Group (distal facies)
- Onverwacht Group (proximal facies)
- Onverwacht Group (distal facies)
- Thrust with teeth on upper plate
- Road
- 6 Tectonic unit

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FIG 2

LEGEND.

Ophiolitic material
- Feldspar porphyry
- Malalotcha Group
- Diepgezet Group
- Lava
- Chert
- Proximal facies spheroids
- Hooggenoeg felsic volcanics
- Distal facies spheroids
- Serpentinite
- Talc schist
- D I tectonite
- Tectonic unit
- Younging directions
- Thrust
- D_2 anticline
- D_2 syncline
- D_2 anticline
- D_3 syncline
- D_3 anticline
- D_3 Fault
- D_4 Fault

Figure Caption: (1) Simplified geological map of part of the investigated area (for location see inset). (2) Three sections, as located on Fig. 1 showing part of the thrust complex. Note how some of the thrust-slices (individually numbered) composed of sediments (Diepgezet and Malalotcha Groups) are tight to isoclinally folded. Folding and thrusting are related to the same regional deformation (D_2).

AGE CONSTRAINTS ON THE EVOLUTION OF THE QUETICO BELT, SUPERIOR PROVINCE, ONTARIO;

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Much attention has been focussed on the nature of Archean tectonic processes and the extent to which they were different from modern rigid-plate tectonics. The Archean Superior Province (1) has linear metavolcanic and metasediment-dominated subprovinces of similar scale to Cenozoic island arc-trench systems of the western Pacific (2), suggesting an origin by accreting arcs (3,4). Models for the evolution of metavolcanic belts in parts of the Superior Province suggest an arc setting (4,5) but the tectonic environment and evolution of the intervening metasedimentary belts are poorly understood. In addition to explaining the setting giving rise to a linear sedimentary basin, models must account for subsequent shortening and high-temperature, low-pressure metamorphism (6-8). Correlation of rock units and events in adjacent metavolcanic and metasedimentary belts is a first step toward understanding large-scale crustal interaction. To this end, zircon geochronology has been applied to metavolcanic belts of the western Superior Province (9-13); this study reports new age data for the Quetico metasedimentary belt, permitting correlation with the adjacent Wabigoon and Wawa metavolcanic subprovinces.

The 10-100 km-wide Quetico belt extends at least 1200 km from beneath cover in the west to the Kapuskasing structure and probably continues 800 km further east, as the Opatica belt. It is mainly fault-bounded against adjacent metavolcanic rocks but stratigraphic contacts are present locally. The belt consists of marginal zones of metasedimentary schist and an interior zone of migmatite and granite. Marginal metasediments have preserved sedimentary structures suggesting a homogeneous sequence of turbiditic greywacke, possibly derived from adjacent volcanic highlands (14). Conglomerate and cross-bedded sandstone of the Seine Group (15) occur sporadically along the northern margin of the belt and have been interpreted as proximal fan deposits of the Quetico turbidites (16) or as a younger sequence (15,17).

The most prominent structural features of the belt are the regular east-trending bedding which dips steeply near the margins and moderately in the interior, and a pervasive, gently east-plunging lineation. Several early sets of folds have been recognized in detailed studies (18-20). Symmetrical low-pressure metamorphic zonation characterizes marginal schists, where grade increases from chlorite-muscovite at the margins, through biotite, staurolite, and garnet-andalusite zones, to garnet-cordierite-sillimanite grade adjacent to the interior zone of migmatite and intrusive granite. Common assemblages of garnet-andalusite throughout marginal schists and locally in the interior indicate low metamorphic pressure (bathozone 2; 3.3 kbar (21)). Granulite facies occurs in the east near Flanders Lake (22) and adjacent to the Kapuskasing zone (23), where metamorphic pressure is 4-6 kbar (24). The regional metamorphic culmination is coincident with interior plutons, suggesting that the granites transmitted heat to high levels in the crust.

Plutonic rocks, classified into three compositional groups, have restricted spatial distribution: 1) a suite of small diorite-monzonite plugs cuts marginal schists and extends locally into adjacent metavolcanic belts; 2) biotite-magnetite leucogranite with local tonalite and amphibolite inclusions, occurs near the schist-migmatite contact; and 3) peraluminous granite, with garnet, cordierite, muscovite, sillimanite, apatite and tourmaline, are prevalent in the interior zone, particularly the Sturgeon Lake batholith (8). Late pegmatites are ubiquitous in the interior zone and common in the higher-grade parts of the marginal schist unit.

QUETICO BELT, SUPERIOR PROVINCE
Percival, J.A. and Sullivan, R.W.

U-Pb zircon geochronology in the Wawa subprovince indicates major volcanic activity between 2749 and 2696 Ma (25) followed by D_1 deformation at about 2696, deposition of alkaline ("Timiskaming") volcanics at 2689, D_2 deformation, and intrusion of post-tectonic plutons at 2684 Ma (9) to 2668 Ma (26) (Fig. 1). In the Wabigoon subprovince, volcanics were erupted in the interval 2755-2702 Ma, with post-tectonic plutons younger than 2695 Ma (12) (Fig. 1).

A chilled porphyritic dacite sill cutting biotite-grade Quetico metasediments yielded an imprecise U-Pb zircon date of 2743 ± 16 Ma, providing a minimum age for sediment deposition. A single tonalite clast from metaconglomerate at Max Creek, interpreted to be Seine equivalent, has zircons dated at 2684 ± 10 Ma, interpreted as the age of the source pluton. Together these dates show that the Quetico metasediments and Seine Group are not facies equivalent. Monazites from the geologically oldest plutonic rock type, a foliated biotite granite with zircons with relict cores, are discordant, with an upper intercept of 2684 Ma. Monazite from massive peraluminous granite with probable inherited zircon is concordant at 2670 Ma. Zircon and monazite from a pegmatite dyke form a discordia line with an upper intercept of 2671 Ma (Fig. 1). The data do not permit definition of the length of time of sediment deposition nor is the thickness of the sequence known; thus inferences on lithospheric thickness (28) cannot be made.

Preliminary synthesis suggests that sediment deposition on extending crust forming the Quetico basin probably occurred during volcanism in adjacent terranes, possibly continuing until volcanism ceased. Closure of the basin during D_1 and/or D_2 events, dated in adjacent belts, led to folding of the sedimentary pile and thickening of the weak crust. Conglomerates were deposited adjacent to marginal transcurrent faults. During subsequent thermal relaxation, partial melts were extracted from lower crustal metasedimentary and tonalitic rocks in a crustal root zone as well as from the mantle. The derived granites and diorites ascended passively to within 10 km of the surface, producing a regional low-pressure aureole in the host schists. A back-arc or inter-arc setting is favoured over an accretionary prism environment for the Quetico sediments because of its symmetry and high-temperature metamorphism which probably occurred in a region of high heat flow.

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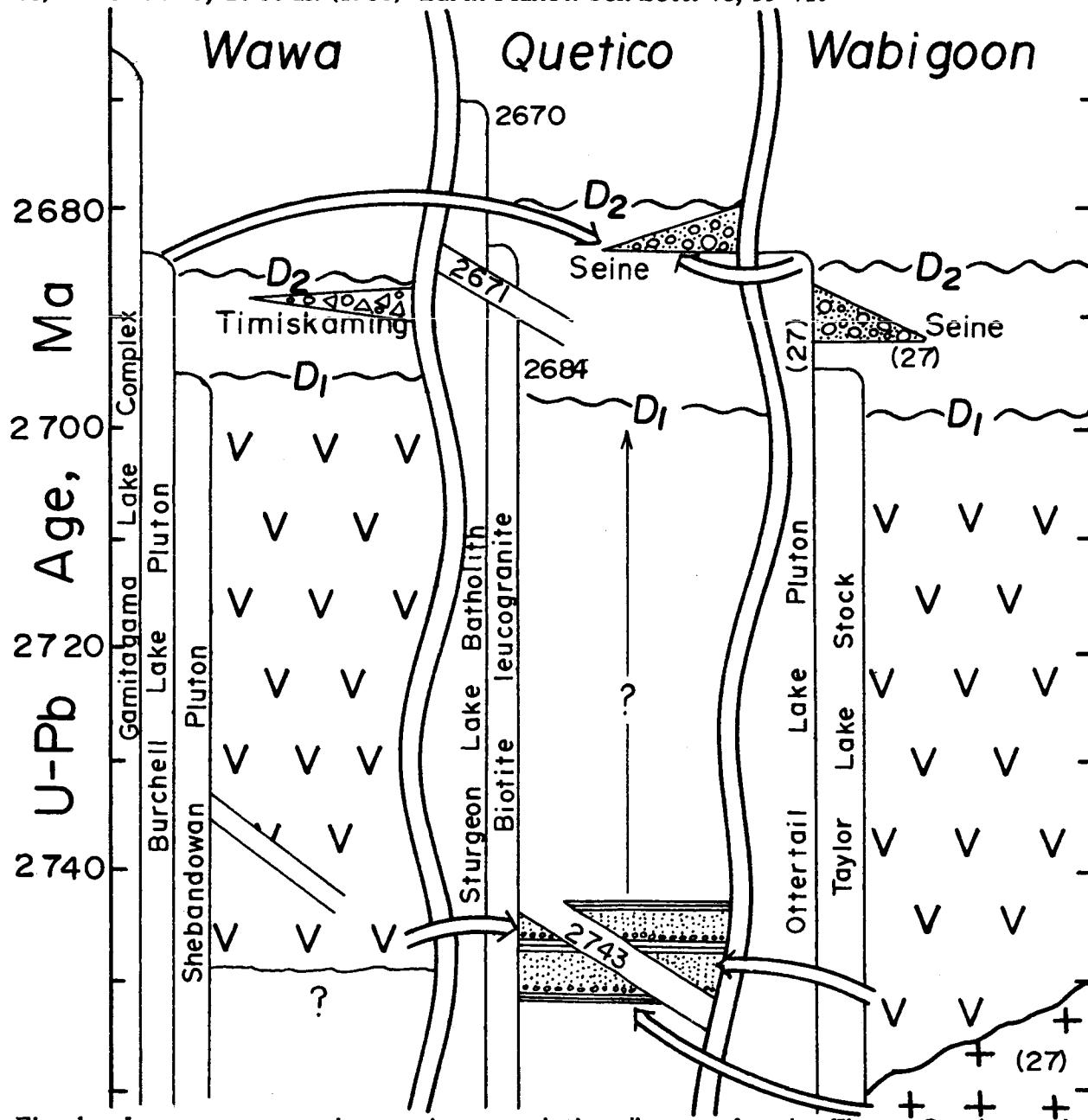
Fig. 1: Age summary and tentative correlation diagram for the Wawa, Quetico and Wabigoon subprovinces. Arrows crossing subprovince boundaries indicate sedimentary provenance.

GREENSTONE BELTS: THEIR BOUNDARIES, SURROUNDING ROCK TERRAINS, AND INTERRELATIONSHIPS

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Introduction

Greenstone belts are an important part of the fragmented record of crustal evolution, representing samples of the magmatic activity that formed much of Earth's crust. Most belts developed rapidly, in less than 100 Ma, leaving large gaps in the geological record. Surrounding terrains provide information on the context of greenstone belts, in terms of their tectonic setting, structural geometry and evolution, associated plutonic activity, and sedimentation.

Tectonic Setting

Major controversy exists as to whether greenstone belts were deposited in oceanic, or marginal oceanic (1-3) or on rifted or thinned sialic crust (4-8). Archean volcanic sequences have much in common with Cenozoic volcanic arcs in terms of linear arrangements, rock types, and sequences, including calc-alkalic volcanic cones built on basal, subaqueous tholeiitic flows. Life spans are 5 to 20 Ma for individual volcanoes and 50 to 100 Ma for individual greenstone belts; some granite-greenstone terrains have several volcano-plutonic cycles differing in age by 200-300 Ma. Associated sediments consist of thin sequences of iron formation, chert, carbonate, and shale, and aprons of immature volcanogenic turbidites. Significant differences include the relative abundance of komatiites, the bimodal nature of some Archean sequences compared to the dominantly andesitic Cenozoic volcanoes, and the paucity of shelf sediments in Archean belts.

Direct evidence of oceanic settings for greenstone belts is rare. A well-preserved ophiolite sequence of Early Proterozoic age is reported from the Kainuu area of Finland (Kontinen, A., written communication, 1985) and a dismembered Archean ophiolite sequence has been interpreted in the southern Wind River Range (9). Neither is evidence for a dominantly continental setting compelling. Although sialic basement to the 2.7 Ga greenstone belts of the Slave and Superior Provinces of Canada has been recognized or inferred at several localities (4,10-13), most granitoid rocks are intrusive into, or in tectonic contact with, the volcanic rocks. Plutonic rocks, commonly with remnants of still-older supracrustal sequences, formed the basement to some volcanic piles, in a continental, micro-continental, or dissected arc setting.

A minor but significant component of Late Archean greenstone belts of the Superior Province is alkaline volcanic rocks, commonly associated with coarse alluvial-fluvial sediments, that unconformably overlie the major volcanic-plutonic successions, only a few Ma older (14-16). These sequences have many similarities to shoshonites formed in recently stabilized arcs (17).

Relationship of Greenstone Belts to Surrounding Terrains

In addition to rare unconformable relationships, fault, intrusive, and conformable depositional contacts characterize greenstone belt margins. Structure within greenstone belts is highly variable in both style and intensity of deformation. Common

features include sinuous, bifurcating folds, steep foliation and lineation and internal shear zones. Deformation may result from several causes, including: 1) tectonic emplacement of the belt (18-21); 2) diapiric rise of external and internal granitoid bodies (18,22-24); and 3) regional compression and/or transpression (25-27). In Slave Province stratigraphic onlap relationships between overlying greywacke-shale sequences and underlying volcanic rocks are common. This contrasts with the Superior Province, where belts of sedimentary rock, fault-bounded for the most part, alternate on a 50-150 km scale with major volcanic-plutonic belts.

As well as discrete fault contacts that form many belt boundaries, complex intercalation of volcanic and plutonic or sedimentary rocks by thrusting has been recognized in widespread locations (19,28-31). Thrusting at infrastructural levels may be an important process in high-grade gneissic terrains (32). Transcurrent displacements of at least several tens of kms have been estimated along some subprovince boundaries in the Superior Province (27,33,34), leading to the suggestion that greenstone and sedimentary subprovinces are accreted blocks. (27, 47, 59)

Plutonic Terrains. Plutonic rocks are particularly abundant in Archean volcano-plutonic terrains where they surround and intrude greenstone belts. Lithologically, these include variably xenolithic tonalite gneiss and more homogeneous bodies ranging from diorite to granite and syenite. Many syn-to post-kinematic plutons were emplaced during early magmatic and late diapiric stages spanning time intervals of ca 20 Ma (35). External plutons are generally similar in composition and age to plutons within belts. Although some plutonic rocks are older than and may represent basement to supracrustal sequences, contacts are generally intrusive or tectonic; precise zircon dating in Superior Province has demonstrated that many tonalite-diorite plutons are coeval with the volcanic hosts (13,36,37). Plutons of granodiorite-granite composition commonly post-date the youngest volcanic rocks and major tectonism by 5-25 Ma. Abbott and Hoffman (38) accounted for voluminous Archean tonalitic magmatism by tapping of low-temperature melts from large volumes of hydrous oceanic lithosphere consumed in shallow subduction zones. The equally voluminous granodiorite-granite magmatism may be the result of lower-crustal melting induced by thickening during collisional or accretionary events. (47).

Plutonic terrains east and west of the Kolar Schist belt have been interpreted as distinct continental fragments, sutured along the schist belt (39). Collisional processes between Precambrian blocks have not been substantiated paleomagnetically (40).

Metasedimentary Belts. Large tracts of metasedimentary rock, predominantly greywacke and shale deposited in turbidite sequences, are distinguished from the iron formation-chert-carbonate-shale successions commonly associated with greenstone belts. Metasedimentary belts, commonly metamorphosed to amphibolite facies gneiss and migmatite, constitute a significant supracrustal component of many Archean terrains, most notably the Slave and Superior Provinces of Canada.

Turbidites make up some 80% of the supracrustal sequences within the Slave Province (70). Deposition of sediments of felsic volcanic and plutonic derivation (41), is thought to be broadly coeval with eruption of marginal volcanic sequences of about 2670 Ma age (10), possibly in response to regional extension (42). The turbidites have alternatively been interpreted (20) as trench-fill deposits in a prograding accretionary complex. Sialic basement of 3 Ga age (43,44), recognized at several locations, has

been variably interpreted as continuous pre-greenstone sialic crust or as microcontinental fragments. Low-pressure regional metamorphism results from the rise of thermal domes (45), possibly associated with the intrusion of plutons.

Three major linear metasedimentary belts separate granite-greenstone terrains of the Superior Province (46,47): the English River, Quetico and Pontiac belts. Although volcanic rocks are rare or absent from the turbiditic sequences, a felsic volcanic (48) or mixed volcanic and plutonic provenance (49) is inferred. Sedimentary sequences are generally in fault contact with adjacent terrains and increase in metamorphic grade from low at the margins to high (migmatite to low-P granulite) in axial regions, where plutons, particularly peraluminous monzogranites, are abundant. It is apparent that these belts developed as elongate sedimentary basins collecting detritus from adjacent volcanic-plutonic highlands and were later subjected to deformation, axial plutonism and high-level metamorphism.

The oldest detrital zircons in metasedimentary belts are commonly derived from ancient terrains either not yet recognized, at great distance from sediment deposition, or destroyed, buried or allochthonous subsequent to the erosional event. Examples include 4.2 Ga zircons in the 3.5 Ga Mt. Narryer quartzite (50), 3.1 Ga zircons in the 2.7 Ga Pontiac belt (51), and 3.8 Ga zircons in the 3.7 Ga Nulliak quartzite (52).

Relationship Between Low and High-Grade Terrains. High-grade terrains form large parts of some Archean cratons and have variable relationships to adjacent greenstone belts. Characterized by upper-amphibolite to granulite-facies metamorphic grade in mainly intrusive rock types, high-grade terrains have been interpreted as either lateral equivalents of greenstone belts, in a different tectonic environment (53,2), or as the deeply-eroded roots to greenstone belts (54). Geobarometry is a useful tool in distinguishing between alternative interpretations in specific areas. Recognition of geological and geophysical criteria of crustal cross-sections (55) may also guide interpretation.

Examples of both lateral and vertical transitions from low to high-grade terrains are documented in the Superior Province. A lateral relationship has been inferred for the high-grade Quetico metasedimentary belt and adjacent low-grade Wabigoon and Wawa metavolcanic-plutonic belts. Volcanic rocks were deposited 2750-2695 Ma ago (13,26). Coeval turbiditic metagreywackes of the Quetico belts, about 2744 Ma old (56) have an axial high-temperature, low pressure zone of schist, migmatite, S-type granites and local granulite (58-60), suggesting a major thermal anomaly at high structural levels. Different tectonic settings and evolution are proposed for the low- grade volcanic (arc) and high-grade metasedimentary (marginal basin) terrains. Differences in structural style between belts can be attributed to variable levels of exposure (60) or mechanical character.

Evidence of dextral transpressional deformation characterizes the Wawa-Quetico-Wabigoon boundary region. This includes: 1) assymetric folds and other kinematic indicators in the northern Wawa (26), Quetico (60) and southern Wabigoon (27) belts, and 2) conglomerate and alkaline volcanic deposits associated with strike-slip faults (27,26). The event is bracketed between 2695 and 2685 Ma by zircon dates (13).

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Adjacent high and low-grade Archean terrains have been interpreted, by analogy with the Cenozoic Rochas Verdes complex (2), as deeply-eroded arcs and adjacent back-arc basins respectively.

Vertical relationships between low and high-grade regions have been interpreted in the intracratonic Kapuskasing uplift (61,62) and marginal Pikwitonei region (63) of the Superior Province, as well as in the Kaapvaal Craton (64). An uninterrupted oblique cross-section through the Michipicoten greenstone belt to lower crustal granulites is exposed across a 120-km-wide transition in the southern Kapuskasing uplift. Well-preserved metavolcanic and metasedimentary rocks of the greenstone belt, metamorphosed to greenschist facies at 2-3 kbar, are intruded and underlain by some 10-15 km of tonalitic rocks which increase in structural complexity from homogeneous plutons to contorted gneisses with increasing depth. Lowermost in the section is a heterogeneous granulite complex, at least 10 km thick, of interlayered supracrustal (15%) and intrusive (85%) rocks recording metamorphic conditions of 700-800°C, 7-8 kbar (66). The crustal slab was emplaced onto low-grade rocks of the Abitibi belt on the Ivanhoe Lake thrust (66) some 2 Ga ago.

In the Pikwitonei region, distinctive rock types including iron formation, pillow basalt, calc-silicates and anorthosite can be traced along strike from the low-grade Sachigo Subprovince into Pikwitonei granulites (63). Supracrustal rocks step up in metamorphic grade across faults (67) as intrusive rocks become more abundant. Metamorphic pressure increases within the granulites from 7 to 12 kbar (68) toward the western boundary, the Nelson Front. Both the Kapuskasing and Pikwitonei structures have diagnostic features of crustal cross-sections including gradients of metamorphic grade and pressure, high proportions of intrusive rock types and paired gravity anomalies.

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ARCHEAN MEGACRYSTIC PLAGIOCLASE UNITS AND THE TECTONIC SETTING OF GREENSTONES
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Large (up to 20 cm), equidimensional, commonly euhedral, plagioclase mega-crysts of highly calcic composition (An_{80-90}) occur commonly in all Archean cratons in one or more of three distinct associations:
1) as cumulate crystal segregations of anorthosite or as megacrysts in basal-tic dikes, sills, and flows in greenstone belts that vary in metamorphic grade from greenschist to granulite. Throughout 100's of thousands of square kilo-meters of northwestern Ontario and Manitoba the plagioclase megacrysts occur in pillowed and massive flows, sills, dikes, large inclusions in dikes, and intrusive anorthositic complexes (Fig. 1) with areas of up to a few 100 km^2 and spanning a period of at least 100 m.y. in the 2.7 to 2.8 b.y. time frame,
2) as basaltic dike swarms in stable cratonic areas forming parallel to sub-parallel patterns over hundreds of thousands of square kilometers intruding both granitic gneisses and supracrustal belts including greenstones. These swarms include the Ameralik-Saglek system at 3.1 to 3.4 b.y. (Fig. 2) [1], the Matachewan system at 2.5 to 2.6 b.y. [2], and the Beartooth-Bighorn system at 2.2 to 2.3 b.y. [3], and
3) as anorthositic complexes associated with marbles and quartzites (Sittam-pundi, India and Messina, South Africa) in granulite grade terrains.

Initial attempts to correlate tectonic settings of similar modern cryst-bearing units with their Archean counterparts were only partially successful. Plagioclase phenocrysts of An_{80-90} occur in basaltic volcanic flows in oceanic crust at spreading ridges, hotspots, aseismic ridges, and fracture zones [4]. These recent occurrences, however, normally involve only small phenocrysts up to a few millimeters in size and usually more lathy than equidimensional in shape [5]. In contrast to these normal occurrences, volcanic flows over the Galapagos hotspot display more equidimensional crysts up to 3 cm across [4]. Although these oceanic environments might be satisfactory tectonic analogs for many greenstone occurrences, they certainly are not satisfactory for the exten-sive dike swarms in stable cratonic masses. Thus we turn for clues to a more detailed understanding of the petrogenesis of the crysts and related melts.

The crysts are quite homogeneous, varying by little more than one to two An units over several centimeters thereby suggesting nearly isothermal crystal-lization at nearly constant melt composition over the time required to grow crystals commonly 6 to 8 cm across and up to 20 cm across and accumulate them in large masses. Thin, more sodic rims on the order of 100 to 200 μm wide are common on large crysts when the groundmass plagioclase laths are more sodic than the large crysts. The rims normally approach the composition of the plagioclase in the groundmass (Table 1).

The nature of the parent melts, or melts in equilibrium with the large crysts, has been an open question because: 1) the anorthositic complexes are clearly cumulates with bulk compositions too rich in Al_2O_3 and CaO to repre-sent melts [6], and 2) the disparity in composition between plagioclase crysts and plagioclase of the matrix suggests a lack of equilibrium between crysts and the melt represented by their matrix.

Initial attempts to determine melt compositions by use of REE concentra-tions in megacrysts in conjunction with distribution coefficients for plagio-clase and basaltic melts were fraught with problems resulting from modifica-tion of plagioclase REE concentrations by alteration, recrystallization, and tiny inclusions. By utilizing several splits from each cryst in several samples from the BVL anorthosite, mixing lines were determined and the least

modified REE concentrations were calculated for pristine plagioclases [7]. These values in conjunction with the most recent distribution coefficients indicate melts with nearly flat REE patterns at 10X to 20X chondrites with perhaps a slight depletion in the light REE's. The calculated patterns compare well with several cryst-bearing basalts in greenstone belts (Fig. 3) as well as with the non-cryst-bearing basalts. These patterns are those of the least enriched tholeiitic basalts which are very common in greenstone belts. Comparison of these basalts with those in the cratonic dike swarms shows many similarities (Fig. 3, Table 2) but the initial data suggests that the cratonic dikes are slightly enriched in SiO_2, K_2O, and light REE. It is tempting to attribute these differences to contamination of the melts as they rise through continental crust but the melts of the Galapagos when compared with MORB show some of the same enrichments (Table 2) which in this case cannot be attributed to continental contaminants. Further work on the pristine REE contents of plagioclase megacrysts is underway and should help determine whether megacrysts in enriched melts formed from the more enriched or less enriched tholeiitic melts, or both.

At present the petrogenetic data require, at a minimum, isothermal crystallization of plagioclase megacrysts from tholeiitic melts (the least enriched ones in greenstone belts) followed by segregation of the plagioclase crystals which then become entrained in rising melts to form intrusions or volcanic flows. Furthermore, the occurrences seem to require large volumes of melt at similar temperatures for long periods of time over huge areas having both oceanic and cratonic associations. Continual generation of similar melt and continuous addition of the melt to extensive networks of crystallizing chambers is also strongly implied. The major remaining questions with significant implications for the setting and evolution of greenstone belts are: 1) Does the cryst-producing melt have the same composition and crystallize under the same conditions beneath greenstone belts, stable cratons, and current oceanic crust? 2) Where do the plagioclase crysts form and accumulate; in low or high pressure environments? 3) Is there a systematic change in the time of megacryst emplacement across large areas such as might be produced by plates overriding zones of melt production or other such time-dependent mechanisms?

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Table 1

Natachewan Dikes: Plag. Compositions

LOCATION	CRYSTS			MATRIX LATHS	
	main cryst	inner rim	outer rim	core	rim
Sta 8	An79-81	An56	An42-49	An61-64	An47
9-11	80-82	66-69	54	65-69	54
10	75-78	61-62	54	58-61	45-50
16	81-84			62-67	68-70	52-53
16	82-84		64-65	64-66	46-53
16	83-85		72-76	71-72	66-69
18	Too Altered			50-54	33
34	81-83			No laths	
35	81-83		78-80	61-63	51-54

Table 2. Matrices of Cryst-Bearing Tholeiites

	RECENT		ARCHEAN	
	MORB Leg 37	Plata Galapagos	Bird River Greenstone	EDM-5 Cratonic Dike
SiO_2	48.36	48.60	48.86	50.00
TiO_2	0.89	2.10	0.96	0.94
Al_2O_3	16.05	16.68	15.00	14.67
Fe_2O_3	3.39	6.52	2.14	2.37
FeO	5.41	3.95	10.14	9.74
MnO	0.17	0.17	0.22	0.19
MgO	7.21	7.23	6.50	6.43
CaO	12.53	12.15	12.07	10.54
Na_2O	2.15	2.80	1.86	2.32
K_2O	0.19	.43	0.29	0.61
P_2O_5	0.08	.31	0.09	0.10
H_2O	2.35	.47	1.76	1.93
La	4.41 ppm	13.00	4.04	9.64
Ce	10.5	36.20	10.12	19.6
Sm	2.11	5.49	2.42	2.96
Eu	0.80	2.02	.895	0.95
Yb	2.19	3.40	2.41	2.414
Lu	.38	.40	.377	0.408

Plagioclase Crysts and Greenstones
Phinney, W. C., Morrison, D. A. and Maczuga, D.
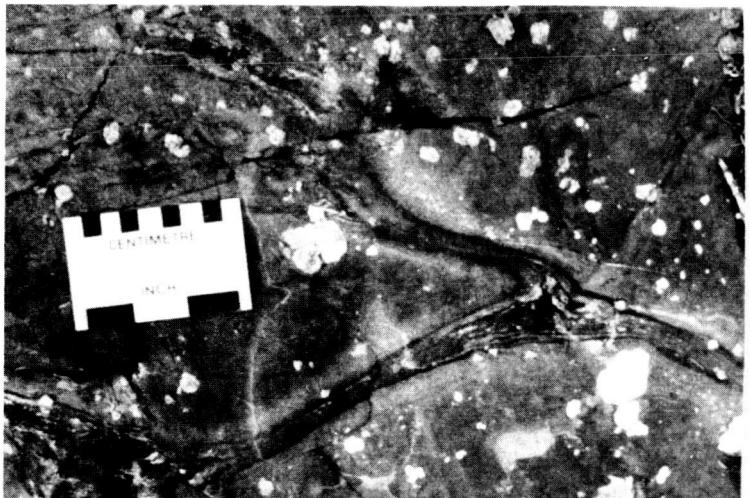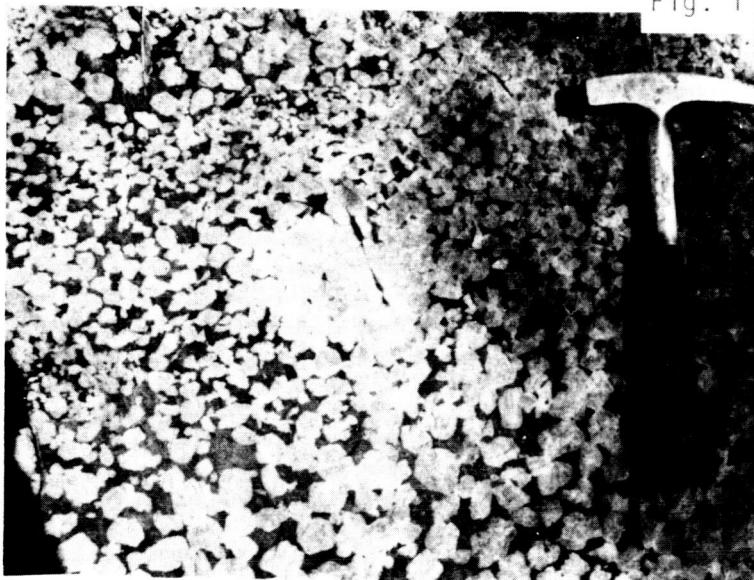
Fig. 1

Fig. 2

BIRD RIVER N. MATRICES   UTIK LAKE MATRICES   MATACHEWAN MATRICES

Fig. 3
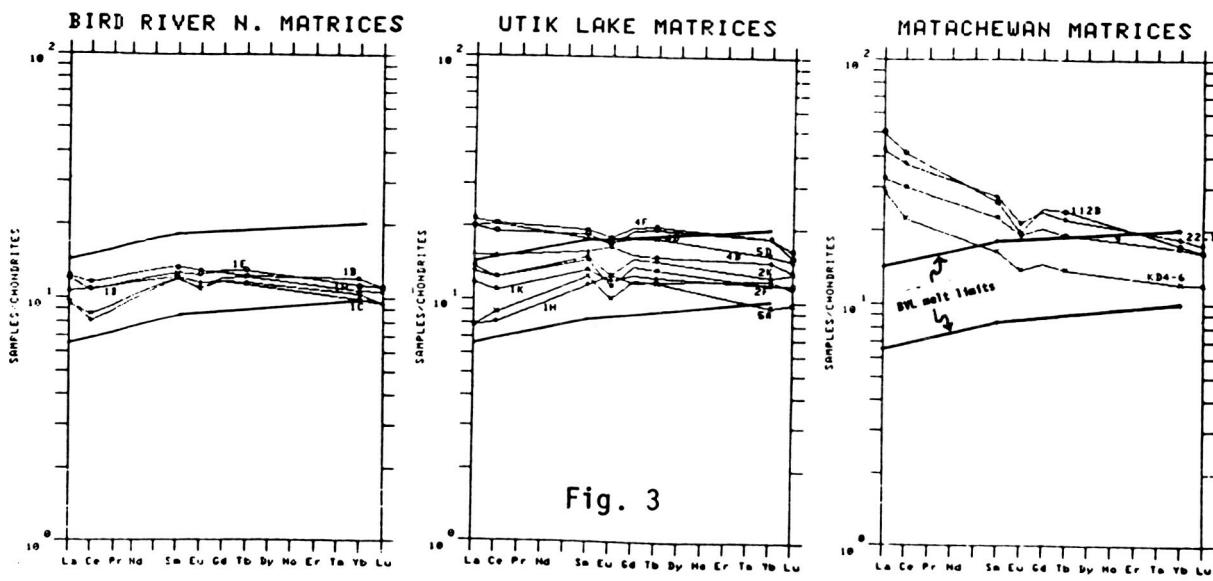
RAINY LAKE WRENCH ZONE: AN EXAMPLE OF AN ARCHEAN SUBPROVINCE
BOUNDARY IN NORTHWESTERN ONTARIO; K. Howard Poulsen, Economic Geology and
Mineralogy Division, Geological Survey of Canada
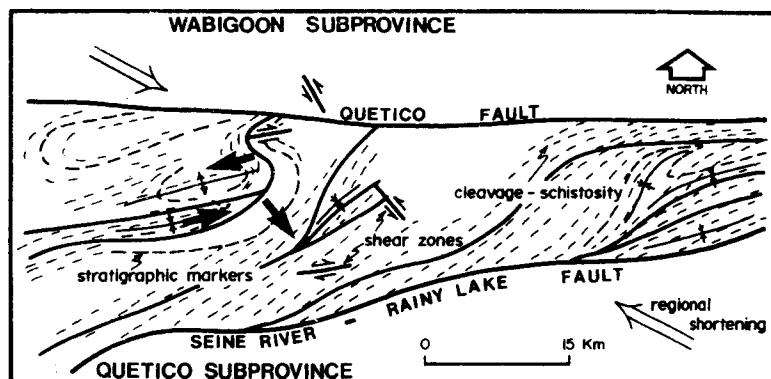
Fig. 1- Schematic diagram
illustrating structural
features of Rainy Lake
Wrench Zone. Short solid
arrows identify downward
facing units.

The Superior Province of the Canadian Shield comprises an alternation of
subprovinces with contrasting lithological, structural and metamorphic
styles (1). Rocks of the Rainy Lake area form a fault bounded wedge between
two of these subprovinces, the Wabigoon granite-greenstone terrain to the
north and the Quetico metasedimentary terrain to the south (Fig. 1). The
Quetico and Seine River-Rainy Lake Faults bound this wedge within which inter-
pretation of the stratigraphy has been historically contentious. In the
eastern part of the wedge, volcanic rocks and coeval tonalitic sills are uncon-
formably overlain by fluviatile conglomerate and arenite of the Seine Group;
in the western part of the wedge, metamorphosed wacke and mudstone of the
Coutchiching Group are cut by granodioritic plutons. The Coutchiching Group
has previously been correlated with the Seine Group and with the turbiditic
Quetico metasediments of the Quetico Subprovince and these correlations are
the cornerstone of earlier tectonic models which relate the subprovinces (2,3).

The structural geology of the Rainy Lake area is characterized by the
following attributes:
(i)   lenticular lithostratigraphic domains with discordant boundaries,
(ii)   steep boundary faults,
(iii) regular orientation and sense of displacement of small ductile shear
       zones,
(iv)   regionally developed sub-vertical foliation which transects large litho-
       logical folds,
(v)   shallow bimodal orientations of minor folds and lineations and a prepon-
       derance of folds of dextral asymmetry,
(vi)   downward facing folds in the Rice Bay, Nickel Lake and Bear Pass areas
       (arrowed, Fig. 1).

These observations compare favourably with the known characteristics of dextral
wrench or "transpressive" zones based both on experimental data and natural
examples (4,5,6,7,8). Much of this deformation involved the Seine Group, the
youngest stratigraphic unit in the area (9), and predates the emplacement of
late-to-post-tectonic granodioritic plutons for which radiometric data indicate
a Late Archean age.

The interpretation of a wrench zone separating the Wabigoon and Quetico
Subprovinces has important implications regarding the tectonic models which can

be used to relate them. Of great importance is the high probability that this zone contains rocks which are actually allochthonous relative to those adjacent in the Quetico and Wabigoon. Given this type of structural environment, not only is correlation of stratigraphic units between individual lenticular domains difficult to establish simply on the basis of some lithological similarity but more important, the correlation with units exterior to the wrench zone is even more suspect. New geochronological data (9) which demonstrates a 40 Ma difference in age between the Seine and Coutchiching strongly supports this argument. Therefore the concept that Seine-type alluvial-fluvial rocks, which are restricted spatially to the wrench zone are transitial "facies" between Wabigoon volcanics and Quetico turbidites (2,3) finds little support in a wrench zone interpretation.

Pettijohn (10) was the first to emphasize that Seine-type sedimentary sequences occur all along the subprovince margin. Because these rocks also correlate spatially with a well defined wrench zone it is instructive to inquire whether an alternate hypothesis might account for these observed relationships without relying on the concept of facies equivalence. The link between alluvial-fluvial sedimentation and wrench zones is well-known in Cenozoic environments where thick alluvial, fluvial and lacustrine sequences are restricted to narrow "pull-apart" basins associated with large transcurrent faults (11,12,13). Such basins are localized by bends in marginal faults and by intersections with fault splays. Lateral and vertical facies variations are present within such basins (14) but these rocks are not contiguous with rocks external to the basin. The size and geometry of the wrench fault system at the southern margin of Wabigoon subprovince and the areal extent of the Seine-type rocks are comparable with younger examples in which there is also a juxtaposition of rocks of differing lithology. In many of these examples, and possibly in the present one as well, the juxtaposed terranes have depositional histories which are quite independent so that present geographic geometry has no simple paleographic significance.

The proposal that wrench faulting is significant at the subprovince boundary is not a new one. Hawley (15) first suggested a model of this type for rocks in the Atikokan area to the east of Rainy Lake but the emphasis in the past has been placed only on the late-stage displacements on the Quetico Fault (2,16) rather than the possibility presented here that such late faulting is merely a reflection of a broader zone of wrenching which also became a locus for sedimentation.

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A PALAEOMAGNETIC PERSPECTIVE OF PRECAMBRIAN TECTONIC STYLES

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The considerable success derived from palaeomagnetic studies of Phanerozoic rocks with respect to the tectonic styles of continental drift (1) and plate tectonics (2), etc. have not been repeated by the many palaeomagnetic studies of Precambrian rocks. This is undoubtedly related to the vast amount of Precambrian time compared with Phanerozoic time, and the concomitant uncertainties of magnetisation ages and rock ages, yet it is still surprising that there is little evidence of consolidation or even convergence of opinions regarding tectonic styles prevalent during the Precambrian. After all, there are 30 years of research with results covering the major continents for Precambrian times that overlap considerably yet there is no concensus even in the grossest terms. There is good evidence that the usual assumptions employed by palaeomagnetism are valid for the Precambrian which only serves to exacerbate the problem. The existence of magnetic reversals during the Precambrian, for instance, is difficult to explain except in terms of a geomagnetic field that was predominantly dipolar in nature. It is a small concession to extend this notion of the Precambrian geomagnetic field to include its alignment with the Earth's spin axis and the other virtues of an axial geocentric dipole that characterise the recent geomagnetic field. In addition it is not a forceful argument to claim that early studies of Precambrian rocks need to be re-done, since re-studies have often only served to confirm the early works. Therefore we submit that the palaeomagnetic results derived from Precambrian rock units are not easily dismissed. It is simply untenable that the majority of the data are spurious and claims that synopses of Precambrian data are invalid, cannot be sustained in such terms. Such arguements posed against the evidence for continental drift have long been debunked. There are, nevertheless, differing interpretations of Precambrian palaeomagnetic data and it is the purpose of this brief article to address this problem.

Methods that have been used to interpret Precambrian palaeomagnetic data fall into two classes. The first class assumes the existence of a "Pangaea" or some supercontinent and proceeds to use the palaeomagnetic data, a posteriori, to support the model. The second class, which we prefer, accepts the palaeomagnetic data at face value (as synthesised by workers closely in touch with the results) and proceeds to view the overall relationships of the data, isolated from preconceived notions. This latter approach has led us to suggest that the present day geographical relationships of continents (from which a reasonable amount of data for the Precambrian are available) yields the more satisfactory comparison. Of course small adjustments of the continents refine this comparison, but overall an excellent agreement in Precambrian pole paths can be realised by leaving the continents in their present locations.

Limitations of the available data in our earlier comparisons (3) restricted the time span of comparisons between different continents to 2300 Ma - 1900 Ma for North America and Africa and 1800 Ma - 1600 Ma for North America, Greenland and Australia. Recently two results have been derived from igeneous rock about 2900 Ma in age, in Australia and Africa. The palaeomagnetic pole positions from these rock units are in close proximity,

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suggesting that the present geographic relationship of Australia and Africa is valid for 2900 Ma ago. The pole position from the Millindinna Complex, Australia, dated at 2860±20 Ma is at 11.9°S, 161.3°E, dp=6.8°, dm=8.4°(4), while the pole position from the Usushwana Complex, Africa, dated at 2880 Ma is at 11.6°S, 165.8°, dp=5.1°, dm=7.5°(5). Thus there is evidence that during the Precambrian North America and Australia were in their present relative geographic locations for 1800 Ma-1600 Ma, as were North America and Africa for 2300 Ma-1900 Ma, and now Africa and Australia, at least for 2900 Ma ago.

These observations are not easily reconciled with Phanerozoic palaeomagnetic results as we have already discussed(3), but they are a matter of record and must be explicable. In terms of greenstone terranes it is obvious that tectonic models postulated to explain these observations are paramount in understanding Precambrian geology. What relevance the current geographical relationships of continents have with their Precambrian relationships remains a paradox, but it would seem that the ensialic model for the development of greenstone terranes is favoured by the Precambrian palaeomagnetic data.

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THE WISCONSIN MAGMATIC TERRANE: AN EARLY PROTEROZOIC GREENSTONE-GRANITE TERRANE FORMED BY PLATE TECTONIC PROCESSES; Klaus J. Schulz, U.S. Geological Survey, Reston, VA 22092 and Gene L. LaBerge, Department of Geology, University of Wisconsin-Oshkosh, Oshkosh, WS 54901 and U.S. Geological Survey

The Wisconsin magmatic terrane (WMT) is an east trending belt of dominantly volcanic-plutonic complexes of Early Proterozoic age.(~1850 m.y.) that lies to the south of the Archean rocks and Early Proterozoic epicratonic sequence (Marquette Range Supergroup) in Michigan. It is separated from the epicratonic Marquette Range Supergroup by the high-angle Niagara fault, is bounded on the south, in central Wisconsin, by Archean gneisses, is truncated on the west by rocks of the Midcontinent rift system, and is intruded on the east by the post-orogenic Wolf River batholith.

Although the history of the WMT is complex in detail, integration of recent studies (Sims and others, in press) provides an overview of its nature and evolution. The WMT shows many similarities to Archean greenstone-granite (AGG) terranes (Condie, 1981). In fact, until recent U/Pb zircon dating, considerable controversy existed as to the age of the rocks of the WMT. Insofar as the comparisons between the WMT and AGG terranes are valid, under-standing of the tectonics of the WMT may provide important insights into the tectonic processes involved in the evolution of at least some AGG terranes.

As in many AGG terranes, a major portion of the WMT is comprised of volcanic rocks and lesser volcanogenic sediments variably metamorphosed to lower greenschist to amphibolite facies. The supracrustal rocks show a complex stratigraphy with at least three successions distinguished on the basis of differences in composition, metamorphism, and structural fabric (LaBerge and Myers, 1984; Sims and others, in press). The older units are dominantly subaqueous basaltic lavas and consanguineous intrusive rocks which are overlain locally by intermediate to felsic volcanic and volcani-clastic units, some in part subaerial (LaBerge and Myers, 1984). Both bimodal (basalt-rhyolite) and calc-alkaline (basaltic andesite through rhyolite) suites are present with the former hosting volcanogenic massive sulfide deposits (May and Schmidt, 1982). Komatiites have not been recognized within the WMT. The older basaltic units are dominantly tholeiitic in character, show strong to moderate depletion of light REE elements ($[La/Yb]_N$=0.09-0.89) and high-field-strength elements (Hf, Zr, Ta, etc.), and are lithologically and compositionally similar to recent back-arc basin basalts (e.g. Mariana Trough, Wood and others, 1981), island-arc tholeiites (e.g. Scotia arc, Hawkesworth and others, 1977), and some ophiolitic basalts (e.g. Troodos, Kay and Senechal, 1976). The younger calc-alkaline units are enriched in LIL elements ($[La/Yb]_N$=2.5-9.5), are also depleted in high-field-strength elements, and are similar to volcanic sequences found in recent island-arcs (e.g. Sunda Arc, Whitford and others, 1979).

Sedimentary rocks are locally found to overlie and(or) interfinger with the volcanic succession. They include graywacke, argillite, thin iron-formation, chert, and minor conglomerate, some containing granitoid boulders (LaBerge and Myers, 1984).

Intrusive rocks within the WMT appear to have been largely diapirically emplaced and show a temporal progression from gabbro and diorite through tonalite and granite. They range from calcic to calc-alkaline in character, although locally slightly alkaline varieties are also present (Sims and others, 1985). The granitoids show an overall increase from north to south across the terrane in their average K_2O/Na_2O ratios and SiO_2 contents. Gneissic rocks, found in domes and block uplifts, are mostly tonalite to granodiroite and are also calc-alkaline (Sims and others, 1985). Both lithologically and chemically, the WMT granitoids appear similar to those formed at compressional plate-margins (Brown, 1982).

Ultramafic rocks are present in the WMT, particularly along the northern and southern margins. They are mostly serpentinized, but perioditic and pyroxenitic lithologies are recognized. These ultramafic rocks are often spatially associated with gabbroic rocks and were in some cases structurally implaced. The ultramafic-gabbroic bodies are lithologically and chemically similar to recent ophiolitic fragments.

Structure within the WMT is complex and consists regionally of large structural blocks having diversely oriented internal structures that are bounded by ductile deformation zones ("shear zones"; LaBerge and Myers, 1984; Sims and others in press). Within the blocks, the supracrustal rocks show generally steep dips and open to isoclinal folds. The deformation zones bounding the blocks record pronounced flattening in the foliation planes and a strong component of verticle movement (Palmer, 1980). This intense deformation along zones is regional in scope, and generally younger than the pervailing internal structural fabric within the blocks. Domes along the northern margin of the terrane, representing large-scale, antiformal fold-interference structures, modified by diapirism and by intrusion of granitoids, have further deformed and metamorphosed the mantling supracrustal rocks (Sims and others, 1985).

U-Th-Pb zircon ages on the volcanic and associated gneissic and granitoid rocks that comprise the WMT (VanSchmus, 1980; Sims and others, in press) indicate that they formed from 1,890 to 1,830 Ma. Detailed isotopic dating in the northeastern portion of the WMT (Sims and others, 1985) indicates that volcanism, granitoid intrusion, metamorphism, and deformation within this region occurred from 1,865 to 1,835 Ma ago, a time span of 30 m.y.

The overall lithologic, geochemical, metallogenic, metamorphic, and deformational characteristics of the WMT are similar to those observed in recent volcanic arc terranes formed at sites of plate convergence. It is concluded that the WMT represents an evolved oceanic island-arc terrane accreated to the Superior craton in the Early Proterozoic. This conclusion is strengthened by the apparent absence of Archean basement from most of the WMT, and the recent recognition of the passive margin character of the epicratonic Marquette Range Supergroup (Larue and Sloss, 1980). On the basis of the new data for the WMT and the epicratonic sequence in Michigan, Schulz and others (in press) have proposed the following tectonic model: 1) early crustal rifting and spreading along the southern margin of the Superior craton, 2) subsequent subduction and formation of a complex volcanic arc, and, 3) with oblique convergence, collision of the arc with the continental margin (epicratonic) sequence and Archean crust of upper Michigan culminating

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in the Penokean orogeny. This tectonic model is similar to plate tectonic histories recently presented for other Early Proterozoic terranes of North America (Hoffman, 1980; Lewry, 1981; Karlstrom and others, 1983). This indicates that the events and processes occurring in the Lake Superior region were not unique, and that the tectonic processes operating were generally similar to those recognized for the Phanerozoic. Given the general similarity of some AGG terranes to the Early Proterozoic magmatic terranes, it seems likely that subduction and plate collisions were also operative in the Archean.

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NEW INSIGHTS INTO TYPICAL ARCHEAN STRUCTURES IN GREENSTONE TERRANES OF WESTERN ONTARIO:

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Ongoing detailed field work in selected granitoid complexes of the western Wabigoon and Wawa Subprovinces, southern Canadian Shield, has led to several new conclusions: (1) Prominent gneiss domes are composed of prestrained tonalite-granodiorite and represent dense hoods of magmatic granitoid diapirs.   The diapiric material commonly was a syenite-diorite crystal mush. (2) The deformation history of the prestrained gneiss remains to be unraveled. (3) The gneiss lacked a thick cover of mafic metavolcanics or other dense rocks at the time of magmatic diaprisim. (4) The synclinoral structure of large greenstone belts is older than the late gneiss domes and may have been initiated by volcano-tectonic processes. Multi-phase granitoid plutonism greatly tightened the synclinoria. (5) Small greenstone masses within the gneiss are complexly deformed, together with the gneiss. (6) No compelling evidence has been found of ductile early thrusting in the gneiss terranes.   Zones   of   greenstone   enclaves   occure   in   hornblende-rich contaminated tonalite and are apt to be deformed magmatic septa. Elsewhere, the tonalite gneiss is biotite-rich and hornblende-poor (FIG 1).

These conclusions rest on several new pieces of structural evidence. (1) Oval plutons of syenite-diorite have magmatic strain fabrics and sharp contacts that are parallel to an axial-plane foliation in the surrounding refolded gneiss. (2) Gneiss domes are lithologically composite and contain large sheath-like structures which are deformed early plutons, distorted earlier gneiss domes, or early ductile nappes produced by folding of planar plutonic septa. (3) The predomal attitudes of gneissosity varied from point to point. It is difficult to prove by conventional structural methods what caused the state of early deformation in the large gneiss domes.   New approaches are being developed based on the patterns of total and incremental finite strain in the granitoid terranes under study.

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NEW INSIGHTS INTO TYPICAL ARCHEAN STRUCTURES
W. M. Schwerdtner
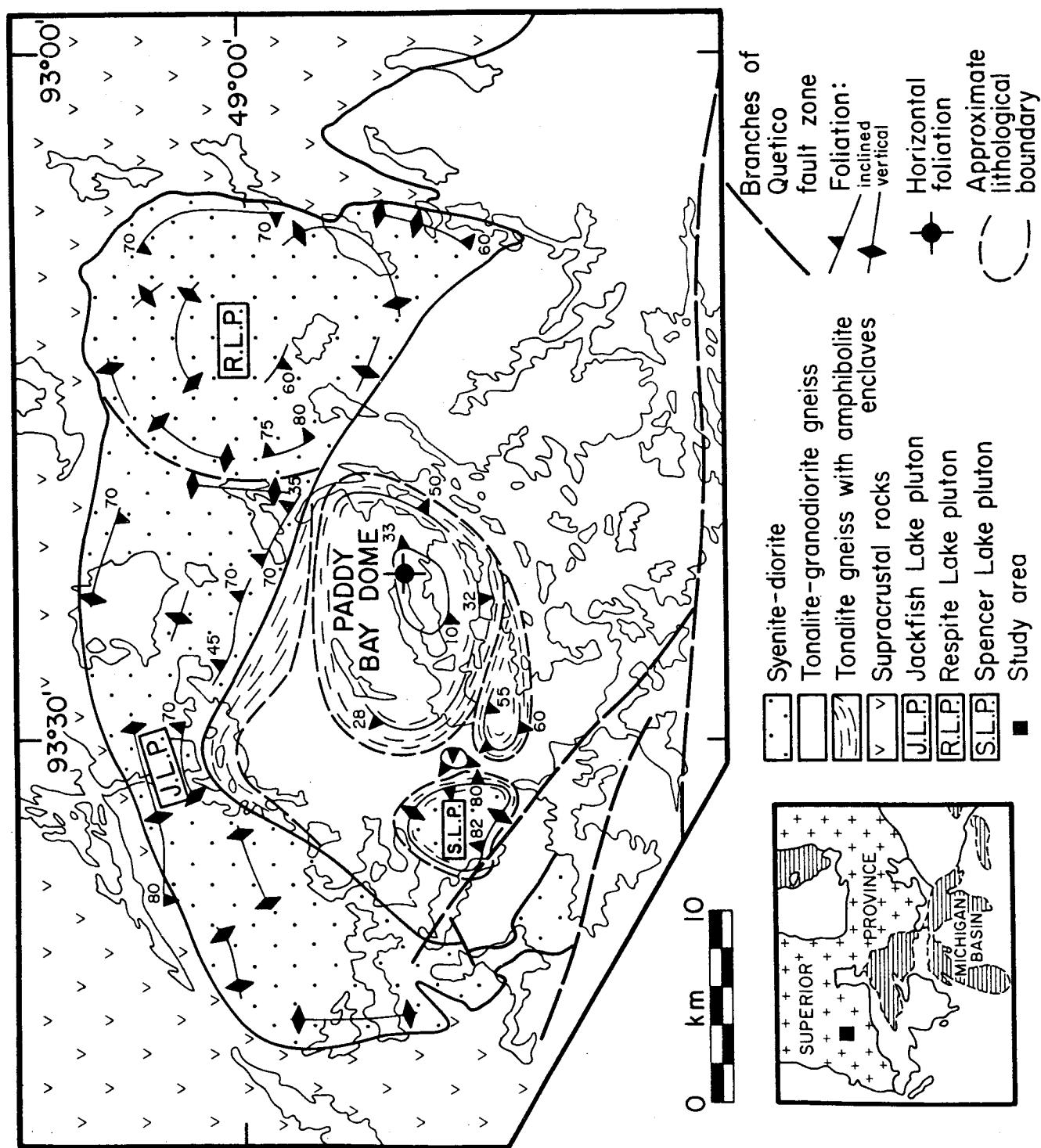
DEFORMATIONAL SEQUENCE OF A PORTION OF THE MICHIPICOTEN
GREENSTONE BELT, CHABANEL TOWNSHIP, ONTARIO; Catherine H. Shrady and
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Detailed mapping at a scale of one inch = 400 feet is being carried
out within a fume kill, having excellent exposure, located in the south-
western portion of the Michipicoten Greenstone Belt near Wawa, Ontario.
A simplified geological map of the area described here is presented
in a companion abstract (Fig. 1 in 1).

The rocks are metasediments and metavolcanics of lower greenschist
facies. U-Pb geochronology indicates that they are at least 2698 ± 11 Ma
old (2). The "lithologic packages" (1) strike northeast to northwest,
but the dominant strike is approximately east-west. Sedimentary struc-
tures and graded bedding are well preserved, aiding in the structural
interpretation of this multiply deformed area.

Deformation in this area is tentatively divided into six phases
(0-5). Phase 0 is soft sediment deformation. Folds of this type are
generally small (amplitudes ranging from several millimeters to tens of
centimeters); however, some early larger scale (up to 10 meters in
amplitude) tight to isoclinal folds with no or a very poorly developed
axial plane cleavage may be slump folds.

Included within Phase 1 of deformation is the regional overturning
resulting in rocks that dip north and young to the south in the northern
part of our area and extending well to the north (1,3,4), and rocks
that dip south and young north in the southern part of our area and
farther south (1,3a). To what extent the regional steep dips are attrib-
utable to this phase of deformation or to later refolding is, at present,
not known. Also included within Phase 1 are an approximately bedding
parallel cleavage, and pebbles within conglomeratic units flattened
parallel to this cleavage. It is thought that these latter two features
are associated and likely relate to the regional overturning.

Cut by and therefore pre-dating Phase 2 cleavage, but of uncertain
temporal relationship to the structures included within Phase 1, are
areally significant faults that separate lithologic packages. These
faults regionally follow but locally truncate bedding. In places,
they are associated with an apparently old fracture cleavage.

Phase 2 is characterized by a penetrative northwest to north strik-
ing cleavage of moderate dip. Phase 2 cleavage crenulates Phase 1
cleavage where both are clearly present; however, in much of the area,
these two cleavages cannot be separated. Related examples of mesoscopic
folds are rare, and associated structures of regional significance
have not been recognized.

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Shrady, C.H. and McGill, G.E.

Phase 3 cleavage is penetrative where well developed and crenulates both Phase 1 and Phase 2 cleavages. Within the area mapped, Phase 3 cleavage strikes northeast with generally steep northwest or southeast dips; dip direction and angle commonly change within individual outcrops. Dips as low as 30° are locally present in the northwest part of the area. It is not clear whether variation in dip indicates the existence of two distinct northeast striking cleavages or whether it is due to later minor folding about sub-horizontal axial surfaces. Phase 3 cleavage is axial planar to folds that are open to tight, range in scale from several millimeters to tens of meters in amplitude, and refold earlier folds. At one locality, Phase 3 cleavage and associated folds appear related to late movement on a fault that approximately parallels bedding. It is not yet clear if this fault is entirely young, or whether it is a reactivated older structure.

Steeply dipping northeast and north-northwest trending faults constitute Phase 4. However, some movement on these faults post-dates diabase dikes (Phase 5) that trend north-northwest and northeast. Locally developed fracture cleavages appear to be associated with diabase dike emplacement, but because the dikes commonly follow trends of older faults, some or all of these fracture cleavages may be related to the faults rather than to the dikes.

In summary: we have tentatively identified at least six phases of deformation within a relatively small area of the Michipicoten Greenstone Belt. These include the following structural features in approximate order of occurrence: 0) soft-sediment structures; 1) regionally over-turned rocks, flattened pebbles, bedding parallel cleavage, and early, approximately bedding parallel faults; 2) northwest to north striking cleavage; 3) northeast striking cleavage and associated folds, and at least some late movement on approximately bedding parallel faults; 4) north-northwest and northeast trending faults; and 5) diabase dikes and associated fracture cleavages. Minor displacement of the diabase dikes occurs on faults that appear to be reactivated older structures.

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A CONTINENTAL RIFT MODEL FOR THE LA GRANDE GREENSTONE BELT;

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Stratigraphic relationships and the geochemistry of volcanic rocks constrain the nature and timing of the tectonic and magmatic processes in the pre-deformational history of the La Grande greenstone belt in the Superior Province of north-central Quebec (Fig. 1). With the exception of a locality in the western part of the belt the lowermost supracrustals in this belt are obscured by syntectonic granitoid intrusives. The supracrustal succession in the western part of the belt consists of a lower sequence of immature clastic sediments and mafic volcanoclastics, overlain by pillowed and massive basalts (Fig. 1, A-A'). Further east, along tectonic strike, a lower sequence of mafic volcanoclastics and immature clastic sediments is overlain by a thick sequence of pillowed and massive basalts, and resedimented coarse clastic sediments and banded iron formation. These are overlain by massive basaltic andesites, andesites and intermediate volcanoclastics intercalated with immature clastic sediments (Fig. 1, B-B'). In contrast, in the eastern part of the belt lenses of felsic volcanics and volcanoclastics occur at the base of the succession and pillowed and massive basalts are overlain by komatiites at the top (Fig. 1, C-C').

The lower sequences of clastic sediments in the central part of the belt reflect a mixed intrabasinal and extrabasinal provenance, but the upper clastic sediments have a uniquely extrabasinal tonalitic provenance. In addition metasedimentary and granitoid xenoliths have been found in the volcanic pile in the central and eastern parts of the belt and a local unconformable contact is believed to exist between the supracrustal succession and an underlying tonalitic basement in the west (1). Therefore a model in which the La Grande belt formed on a sialic crust is favoured.

The largest volumes of eruptive rocks in the La Grande belt are tholeiitic basalts (Fig. 2). These basalts are not primary mantle-derived liquids, but have undergone a polybaric fractionation history (1, 2 and 3). Their parental magmas are believed to have been basaltic komatiites (Fig. 2). The basaltic komatiites and most magnesian basalts lie along a steep slope in Al-Si space (Fig. 2) which is best explained by the fractional crystallization of orthopyroxene and olivine (4, 1). Co-existence of these two silicate phases and a liquid of basaltic composition is restricted
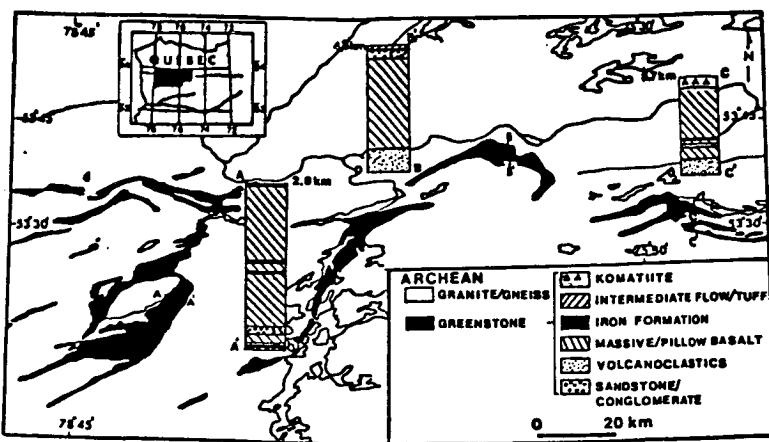

Figure 1 Geology of the La Grande greenstone belt.
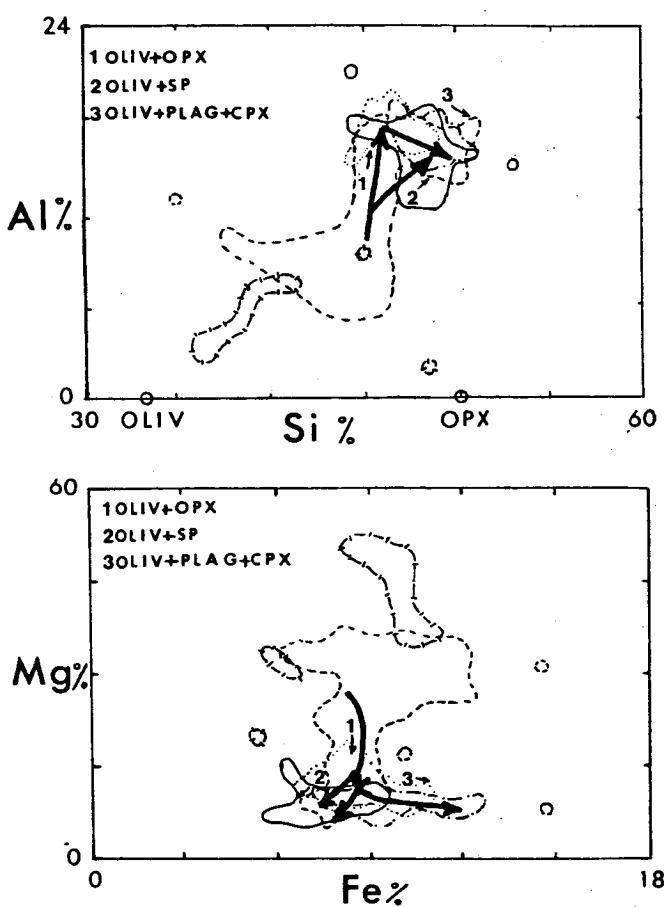
Figure 2 Al-Si and Mg-Fe in cation%. The solid line encloses basalts from section A-A', dotted line is basalts from section B-B', dash-bar and dash-dot are komatiites and basalts respectively from section C-C' and the dashed line includes komatiites and basalts from Lac Guyer (north of C-C).

to pressures on the order of 10 kb (5). Thus the basalts represent komatiitic liquids which have been modified by differing extents of fractionation at depths on the order of 30 km before migrating to higher levels in the crust (3, 1 and 6). A spectrum of basaltic compositions are found in the La Grande belt of which the endmembers are an Fe-enriched suite and those which have negligible Fe variation (Fig. 2). The Fe-enriched basalts have undergone extensive low pressure fractionation of a gabbroic assemblage, which is probably the result of a more protracted residence time in upper crustal conduit system than the relatively constant Fe group. The degree of fractionation of the komatiitic liquids and their location in space and time may reflect the variable efficiency of a crustal density filter (cf. 7). Thus, the occurrence of komatiitic lavas in the upper levels of the supracrustal succession may be due to late failure of the crustal barrier. Their restricion to the eastern parts of the belt may reflect development of a major rift only there. Ponding of mafic magmas within the sialic crust may have resulted in the melting of the crust and the early eruption

of rhyolitic magmas in the east (4). Toward the central parts of the belt, komatiitic magmas ingested sialic crust, were modified by fractional crystallization and were ultimately erupted as basaltic andesites and andesites. These contaminated magmas are characterized by high compatible element (eg. Ni and Cr) and fractionated, enriched light rare earth element abundances (up to 100X chondrite) (8).

The La Grande greenstone belt can be explained as the product of continental rifting (6). The restricted occurrence of komatiites, and eastwardly directed paleocurrents in clastic sediments in the central part of the belt are consistent with rifting commencing in the east and propagating westward with time (Fig. 3). The increase in depth of emplacement and deposition with time of the lower three units (Fig 1, section B-B') in the central part of the belt reflects deposition in a subsiding basin (6). These supracrustal rocks are believed to represent the initial rift succession (c.f. 9). Model calculations (Fig.3) reveal that the extension factor for lithosphere neccessary to account for the observed initial subsidence in the

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central part of the belt (6) is comparable in magnitude with that measured in
Modern sedimentary basins where the continental lithosphere is believed to
have been rapidly thinned (10). The occurrence of clastic sediments of
granitic provenance high in the succession in the central parts of the belt
may reflect the uplift and erosion of marginal forebulges that formed as a
result of lithospheric flexure.
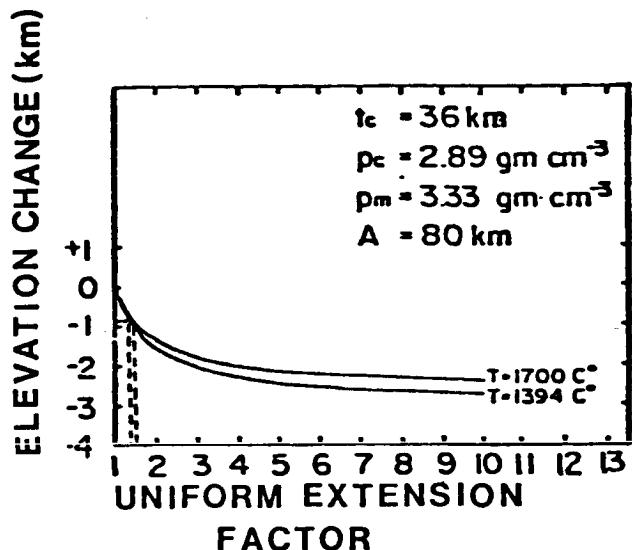
Figure 3 Initial elevation change
versus uniform extension factor. For
an initial elevation change of .9 km
corresponding to the subsidence that
is observed in the lower three units
of section B-B'corrected for the
basin fill and 1 km of water requires
a uniform extension factor of
approximately 1.5. The symbols used
are: crustal thickness (tc), crustal
and mantle densities (pc) and (pm)
respectively, temperature at the base
of the slab (T) and lithosphere
thickness (A).The thermal expansion
coefficient used is 3.2×10^{-5} C. The
calculations were performed using the
method of Royden and Keen (11).

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TWO CONTRASTING METAMORPHOSED ULTRAMAFIC-MAFIC COMPLEXES FROM GREENSTONE BELTS, THE NORTHERN KAAPVAAL CRATON, AND THEIR SIGNIFICANCE IN ARCHAEAN TECTONICS.

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The character of Archaean ultramafic-mafic complexes can, given their prominance in greenstone belts, provide critical clues to help deduce the tectonic setting of these belts. Here we describe two contrasting, metamorphosed, ultramafic-mafic complexes, the first a partially serpentinised dunitic body with associated chromite from Lemoenfontein, one of several peridotitic bodies occuring as discrete lenses and pods in granulite facies gneisses of the northern Kaapvaal craton. The second, the Rooiwater complex is a major layered igneous body, now metamorphosed in the amphibolite facies, but without pervasive deformation, which crops out in the northern Murchison greenstone belt.

The Lemoenfontein body is circular, about 350m in diameter, having the form of a steeply plunging boudin which complements the regional structural pattern. The surrounding granulite facies gneisses were isotopically reset about 2650Ma and may be considerably older. The Lemoenfontein rocks are partially serpentinised dunite, displaying a prominent tectonic fabric defined by the preferred orientation of olivine grains, chromite pods and disseminated chromite stringers, all of which are believed to have been through the granulite facies metamorphism. Chromite is present as massive high-grade ore, 'leopard' (nodular) ore, tectonically layered ore and disseminated ore. Zones of chromite enrichment range in thickness from 1 to 30cm. The Lemoenfontein chromites are similar to those mined in the Ultramafic Formation of the Selukwe greenstone belt, Zimbabwe.

Olivines from Lemoenfontein are Fo94 to Fo96 with NiO contents from 0.35 to 0.59wt%. The mineral chemistry of the chromites of all different types (pods, trains and inclusions in silicate grains) is very similar indicating either complete metamorphic equilibration or they represent consistent primary compositions. The Lemoenfontein chromites have refractory characteristics (low TiO_2, Al_2O_3 and alkali metals) and plot on geochemical fence diagrams in or close to the fields of other podiform chromites. Rocks which in Phanerozoic series are closely associated with alpine-type peridotites or ophiolite suites.

The Rooiwater complex is a thick on end differentiated igneous body, of age greater than 2650Ma, probably intruded at 2960Ma. The complex is heterogenously deformed with much of the 7.5km exposed thickness showing no pervasive deformation. Metamorphosed pyroxenite, anorthosite, gabbro, sulphide-bearing gabbros, thick magnetitite layers and differentiated granites are compatible with the hypothesis that the body is a layered intrusion although it is now allochthonous and intruded by younger unrelated granites. Southward increasing TiO_2 and decreasing V_2O_5 contents in magnetitite layers combined with a

general southerly disposition of differentiated hornblende
granite suggest that the Rooiwater complex is southward facing. A
paucity of ultramafic cumulates and up to 1.5km of highly
differentiated hornblende granite suggests that the original
magma was more felsic than that of similar layered intrusions.

    The Lemoenfontein chromites and associated ultramafic rocks
are lithologically and chemically similar to their Phanerozoic
equivalents of ophiolitic origin, interpreted as obducted oceanic
crust. Similarly we interpret the Lemoenfontein complex as being
a remnant of Archaean oceanic material. In contrast, the
Rooiwater complex is, despite the lack of exposed intrusive
contacts, similar to layered igneous complexes such as Ushushwana
or Bushveld. These complexes are intrusive in continental
environments. We conclude that contrasting ultramafic-mafic
complexes represent a heterogeneity in greenstone belts with
either oceanic or continental environments involved. Whether this
heterogeneity relates to a temporal or spatial (or both) control
remains uncertain.

Figure 1. Geological map of the northern Kaapvaal craton showing
the location of the Lemoenfontein chromites and the Rooiwater
igneous complex.
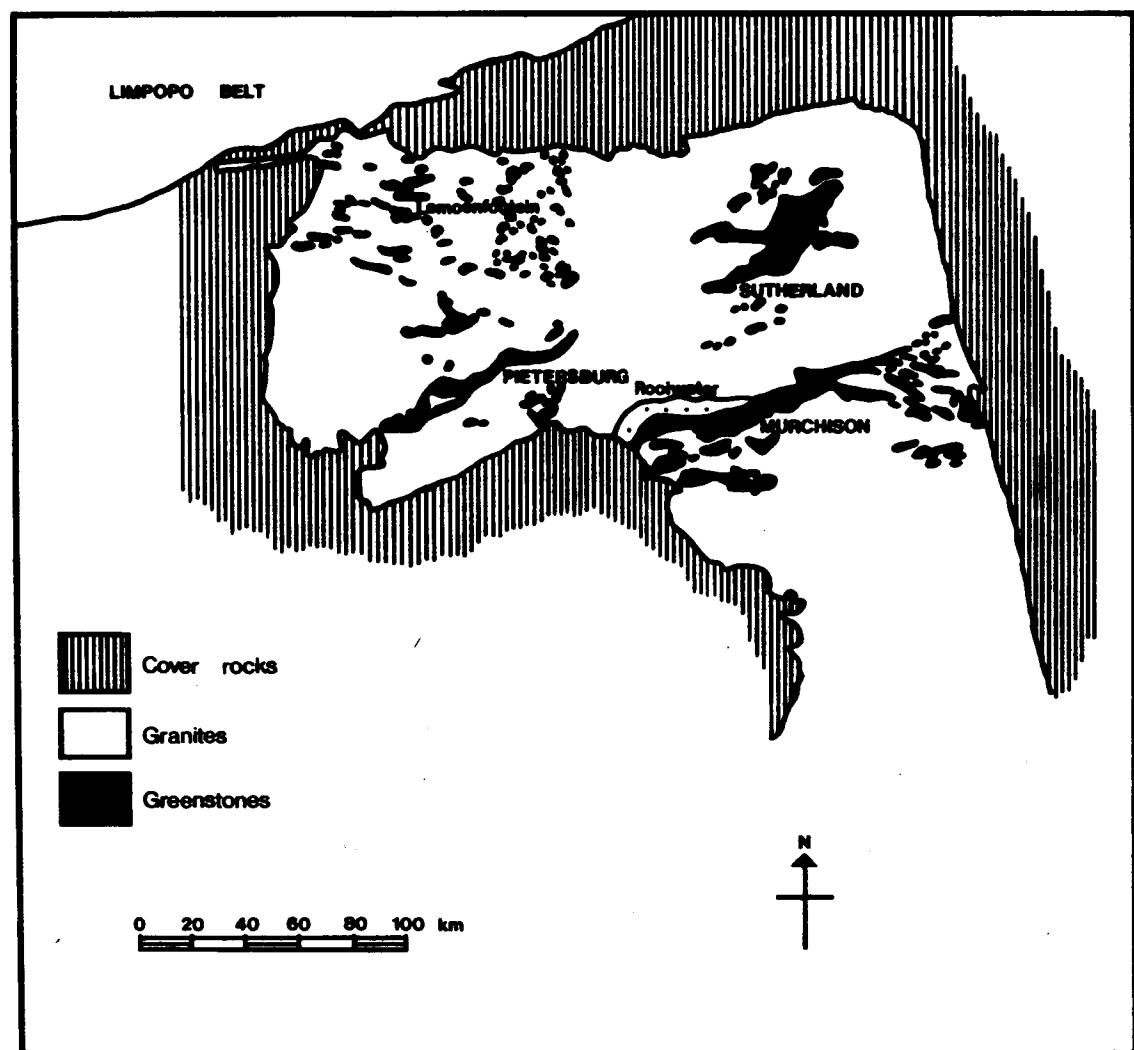
ZIRCON Lu-Hf SYSTEMATICS: EVIDENCE FOR THE EPISODIC
DEVELOPMENT OF ARCHEAN GREENSTONE BELTS P. E. Smith, M. Tatsumoto* and R. M.
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A combined U-Th-Pb and Lu-Hf isotopic study of zircons was undertaken
in order to determine the provenance and age of an Archean granite-greenstone
terrain and to test the detailed application of the Lu-Hf system in various
Archean zircons.

The eastern Wawa subprovince of the Superior province consists of the
low grade Michipicoten and Gamitagama greenstone belts and the granitic
terrain. Earlier studies have established the structural and stratigraphic
relationships of the area (1-4). The adjacent high grade Kapuskasing zone
is believed to represent the lower crustal levels to the greenstone belts (5).

The rock units of this area have been the subject of extensive geochron-
ological studies using zircon U-Pb (6, 7) and whole rock U-Th-Pb methods
(Smith, et al in prep.). The three volcanic cycles recognized in the area
have mean ages of 2748 My (cycle I), 2732 My (cycle II), and 2714 My (cycle
III). Syntectonic granitoids which surround the supracrustal rocks date
from the cessation of cycle I volcanic rocks, to the time of post-tectonic
plutonism dated at 2666 + 2 My. The oldest rocks yet dated come from a
granite dated at 2888 + My which is possibly the basement to the volcanic
rocks. Zircon ages from the Kapuskasing zone appear to reflect updating
during the regional metamorphism (8).

The Lu, Hf, U and Th contents of zircons from these rocks reveal patterns
that may be indicative of their source regions (Fig. 1). Zircons from rocks
of granitic composition appear to have distinct enrichments in U and Th
relative to zircons from rocks of more intermediate composition. More
striking however, is the severe depletion of Lu and Hf from the zircons from
the Kapuskasing area. The lowest Hf content measured so far, 1790 ppm, is
from zircons from a mafic gneiss. The elemental patterns in the lower
crustal zircons suggest that Lu and Hf loss accompanies Pb loss during high
grade metamorphism.

The U-Pb age corrected Hf isotopic ratios from the zircons indicate
significant long-lived heterogeneity of source regions for the greenstone
belts (Fig. 2). Overall the heterogeneity in the ratios may be attributed to
three isotopically distinct sources: (1) a high Lu/Hf source; (2) a moder-
ately enriched Lu/Hf source; and (3) a sub-chondritic Lu/Hf source.

The high Lu/Hf source is represented by a sub-volcanic intrusive from
cycle II and two tholeiites (whole rock determinations) from the lower
stratigraphic levels of cycles I and II. The epsilon Hf values range from
+8.7 to +11.6 and the source is believed to represent the depleted mantle.

The second source has epsilon Hf values ranging from +1.4 to +5.9.
There is an apparent alignment of dacitic volcanic rocks and their sub-
volcanic equivalents from cycles I and II with the tonalitic syntectonic
granitoids. It is believed that the source of these rocks was the lower
crust and it can be inferred that previous intracrustal differentiation led
to a high Lu/Hf lower crustal reservoir. The process which led to the
enhanced Lu/Hf ratio was most likely Hf loss as attested to by the Kapuska-
sing zircons. A greater than chondritic Lu/Hf ratio for the lower crust may
explain the apparent non-coherence of initial Nd and initial Hf ratios for
an Archean tonalite reported in the literature (9, 10).

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The low Lu/Hf source is represented by rhyolites capping the sequences of cycles I and III and by post-tectonic potassic granitoids. Their epsilon Hf values ranging from -1.3 to +1.4, significantly lower than the coeval dacites, are indicative of an upper crustal source.

The Hf isotopic data from the three volcanic cycles indicate that the typical lithological features of a greenstone belt cycle could be accommodated in a crustal growth model that involved decreasing depth of melting in three isotopically distinct reservoirs: mantle, lower crust and upper crust. The model age of the sources given by the intersection of the lower crustal curve with the bulk earth evolution curve (11) is about 2900 My, in good agreement with the zircon U-Pb basement age. This linear array also has a similar inter-section age to that of Proterozoic carbonatite complexes studied by Bell et al (12). The general convergence of the other reservoir vectors around this age suggests that mantle depletion, crustal extraction and intracrustal differentiation were all part of the same episodic event. It is also apparent that recycling of older basement was important in the formation of many of the later greenstone belt rocks.
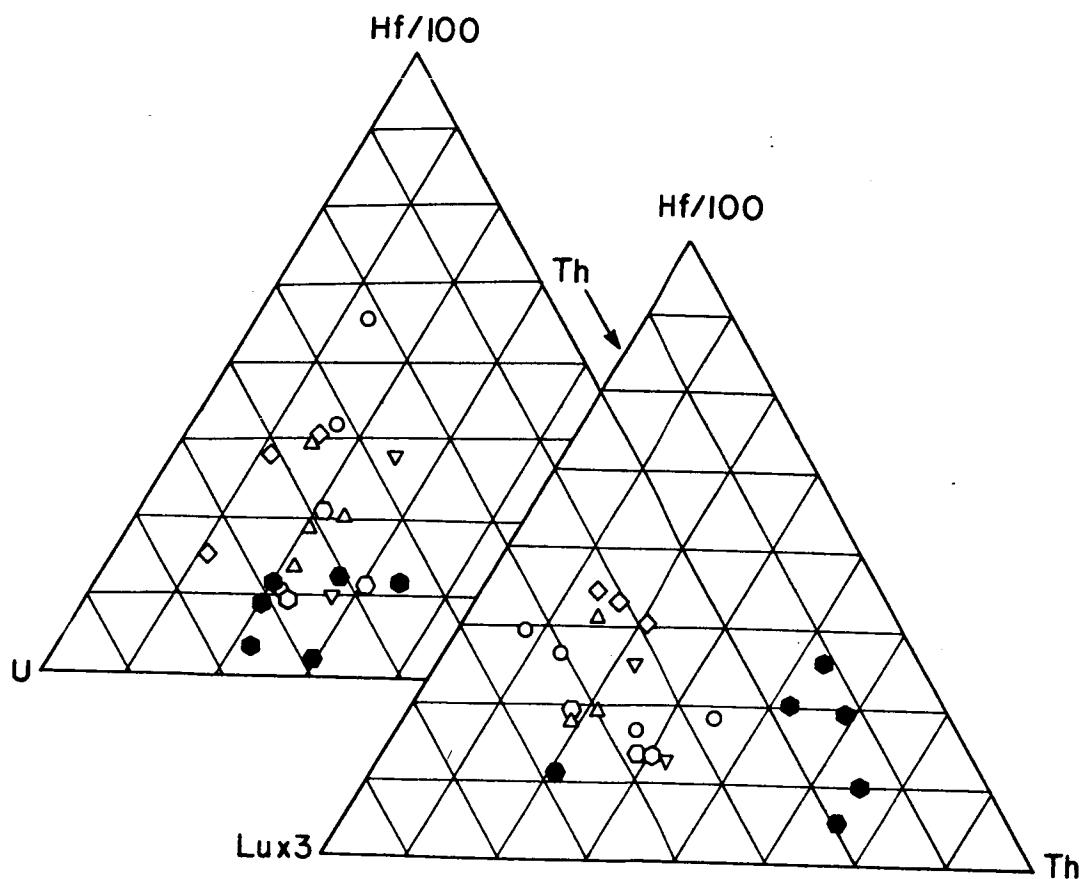

Figure 1

Relative abundances of Lu, Hf, U and Th for eastern Wawa subprovince zircons. Symbols are: △ Dacitic volcanic rocks; ▽ rhyolites; ○ sub-volcanic granitoids; ◇ syntectonic granitoids; ○ post-tectonic granitoids; ○ basement granite zircons; ⬢ Kapuskasing zircons; ◆ conglomerate boulder zircons.

ZIRCON Lu-Hf SYSTEMATICS

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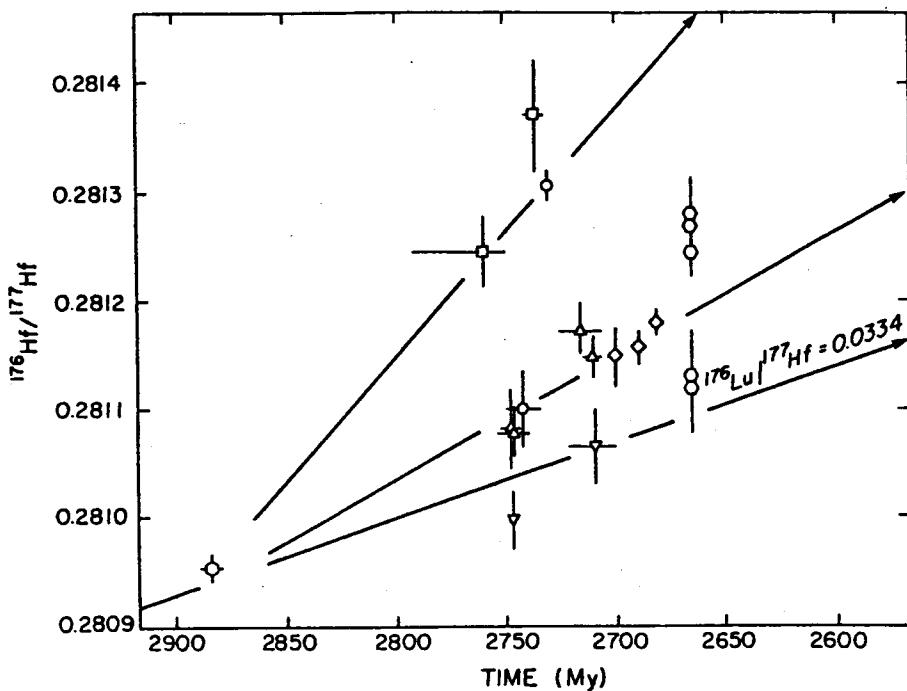
Figure 2

Initial $^{176}Hf/^{177}Hf$ vs T diagram for zircons and whole rocks (☐). Symbols as in Figure 1.

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EARLY PRECAMBRIAN CRUSTAL EVOLUTION OF SOUTH INDIA

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The Early Precambrian sequence in Karnataka, South India provides evidences for a distinct trend of evolution which differs from trends exhibited in many other Early Precambrian regions of the world. The supracrustal rock associations preserved in greenstone belts and as inclusions in gneisses and granulites suggest the evolution of the terrain from a stable to a mobile regime. The stable regime is represented by 1. layered ultramafic-mafic complexes, 2. orthoquartzite-basalt-rhyodacite-iron formation, and 3. ortho-quartzite-carbonate-Mn-Fe formation. The mobile regime which can be shown on sedimentological grounds to have succeeded the stable regime witnessed accumulation of a greywacke-pillow basalt-dacite-rhyolite-iron formation association. Detrital sediments of the stable zone accumulated dominantly in fluvial environment and the associated volcanics are subaerial. The volcanics of the stable regime are tholeiites derived from a zirconium and LREE-enriched source. The greywackes of the mobile regime are turbidites, and the volcanic rocks possess continental margin (island-arc or back-arc) affinity; they show a LREE-depleted to slightly LREE-enriched pattern. The evolution from a stable to a mobile regime is in contrast to the trend seen in most other regions of the world, where an early dominantly volcanic association of a mobile regime gives way upward in the sequence to sediments characteristic of a stable regime.

Structures in greenstone belts, in the gneisses surrounding them, and also in the inclusions in the gneisses are similar in style, sequence, and orientation. This structural unity which is present in spite of the three thermal peaks recorded by radiometric ages around 3300, 3000 and 2600 m.y. ago, indicates long range stability of tectonic stress regimes in the Archaean lithosphere. The continuation of structures and rock formations across the greenstone-granulite boundary suggests that the two provinces did not evolve in separate tectonic blocks but represent only different crustal levels.

The preservation of detrital pyrite-uraninite bearing conglomerates, iron formations, and carbonate rocks provide an unique opportunity for the elucidation of evolutionary changes from oxygen-deficient to oxygenic atmosphere-hydrosphere conditions. Large scale development of iron formations and limestones in the greenstone belts of South India at least 3000 m.y. ago suggests that these may be the earliest large-scale sinks for the photosynthetically produced oxygen. Detailed palaeobiological and biogeochemical studies of these rock formations are necessary.

BASEMENT-COVER RELATIONS AND INTERNAL STRUCTURE OF THE CAPE SMITH KLIPPE: A 1.9 Ga GREENSTONE BELT IN NORTHERN QUEBEC, CANADA;

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The Cape Smith Belt is a 380x60 km tectonic klippe (1 and references therein) composed of greenschist- to amphibolite-grade mafic and komatiitic lava flows and fine-grained quartzose sediment, intruded by minor syn- to post-tectonic granitoids. Previously studied transects in areas of relatively high structural level show that the belt is constructed of seven or more north-dipping thrust sheets which verge toward the Superior Province (Archean) foreland in the south and away from an Archean basement massif (Kovik Antiform) external to the Trans-Hudson Orogen (Early Proterozoic) in the north. A field project (mapping and structural-stratigraphic-metamorphic studies) directed by MRS was begun in 1985 aimed at the structurally deeper levels of the belt and underlying basement, which are superbly exposed in oblique cross-section (12 km minimum structural relief) at the west-plunging eastern end of the belt. Mapping now complete of the eastern end of the belt confirms that all of the metavolcanic and most of the metasedimentary rocks are allochthonous with respect to the Archean basement, and that the thrusts must have been rooted north of Kovik Antiform. The main findings (2) are:

1. A thin autochthonous to parautochthonous low-strain sedimentary sequence on the south margin of the belt rests directly on Archean basement showing no evidence of Proterozoic transposition.

2. The bulk of the belt is separated from the autochthon by a sole thrust which, except at the south margin of the belt, is located at the basement-cover contact. The hangingwall and footwall rocks of the sole thrust record high ductile strains over a zone of increasing width, from south to north, toward the hinterland. Late syn-metamorphic thrusts faults with relatively small displacements cut the sole thrust and its associated shear zone, and place basement gneisses over cover rocks.

3. Lensoid meta-ultramafic tectonic blocks occur locally within the basal shear zone. Their metamorphic anthophyllite-actinolite assemblage differs from the serpentine-tremolite assemblage of cumulate meta-ultramafics occurring in sills at higher structural levels. The blocks may have been tectonically transported from mantle depths during thrusting, although this idea remains to be tested.

4. The allochthonous rocks above the sole thrust occur in a series of thrust sheets bounded by south-verging (D1) thrust faults, which are defined by structural repetitions of stratigraphy and splay from the sole thrust. Favorable lithologies at all structural levels (excepting the southern autochthonous margin) have a pevetrative syn-metamorphic schistosity (S1) which is planar to south-facing tight to isoclinal folds of bedding (F1).

5. A transverse stretching lineation (L1) common in the lower stgructural levels and pervasive in the basal shear zone, when considered with the F1 fold asymmetry and overall thrust-ramp geometry, indicates relative southward translation of the cover during D1.

BASEMENT-COVER RELATIONS AND STRUCTURE OF THE CAPE SMITH KLIPPE
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6. A pelitic interval above the sole thrust on the north margin of the belt contains the metamorphic assemblage kyanite-staurolite-garnet-biotite-muscovite-plagioclase-quartz. The asemblage is indicative of metamorphic T of 550°C and minimum P of 5.5 Kbars.

7. Mesoscopic late- to post-metamorphic chevron to rounded parallel folds (F2) of the S1 fabric have a marked limb asymmetry suggestive of a gravitational origin as folds cascading off basement-cored macroscopic D2 antiforms into pinched cover-rock synforms. The distribution of north- versus south-vergent mesoscopic folds however is not always consistent with the mapped limbs of the macroscopic folds, possibly reflecting diachronous development of the macroscopic folds.

8. Macroscopic high-angle D3 crossfolds affect both the basement and cover in the eastern half of the belt and provide a cumulative structural relief of 12-15 km. D3 fold hinges are readily documented by reversals in plunge azimuth of the D2 folds. Plunge projections permit the construction of a composite structural cross-section linking the highest and lowest structural levels of the belt.

The main implication of these observations is that the presence of Archean basement beneath the belt has no direct bearing on the question of the tectonic setting of the mafic-ultramafic magmatism.

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RHYOLITIC COMPONENTS OF THE MICHIPICOTEN GREENSTONE BELT, ONTARIO:
EVIDENCE FOR LATE ARCHEAN INTRACONTINENTAL RIFTS OR CONVERGENT PLATE MARGINS
IN THE CANADIAN SHIELD? Paul J. Sylvester, SN4, NASA/Johnson Space Center,
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Rhyolitic rocks often are the dominant felsic end member of the bimodal
volcanic suites that characterize many late Archean greenstone belts of the
Canadian Shield [1]. The rhyolites primarily are pyroclastic flows (ash flow
tuffs) emplaced following plinian eruptions [2], although deposits formed by
lava flows and phreatomagmatic eruptions also are present. Based both on
measured tectono-stratigraphic sections and provenance studies of greenstone
belt sedimentary sequences [3], the rhyolites are believed to have been equal
in abundance to associated basaltic rocks.

In many recent discussions of the tectonic setting of late Archean
Canadian greenstone belts, rhyolites have been interpreted as products of
intracontinental rifting [2,4]. A study of the tectono-stratigraphic
relationships, rock associations and chemical characteristics of the
particularly well-exposed late Archean rhyolites of the Michipicoten
greenstone belt, Ontario (figure 1) suggests that convergent plate margin
models are more appropriate.

Three time-equivalent stratigraphic sequences of volcanism (figure 2),
each including both mafic and felsic rocks, have been recognized in the
Michipicoten greenstone belt [5,6,7,8]. The lower volcanic sequence is most
well-preserved and therefore has been studied in most detail. It consists of
a largely mafic unit (MV1) conformably overlain by a thick (up to about
700m), mainly felsic volcanic succession (FV1), which was emplaced
approximately 2743 Ma ago [9]. In the Michipicoten Harbour area, an undated
basal felsic flow unit is structurally discontinuous with the mafic sequence.
Along the northern margin of the belt, epiclastic sediments are deposited on
apparently older granitoid basement, and are overlain by felsic volcanics
(and iron formation) that may be time-correlative with the Michipicoten
Harbour felsic flows.

Fig. 1

Mafic to Felsic Volcanic Rocks
Felsic Intrusive Rocks
Breccia Unit
Doré Formation
Tonalitic to Granitic Gneiss
Facing Direction
Sample Location
Fault
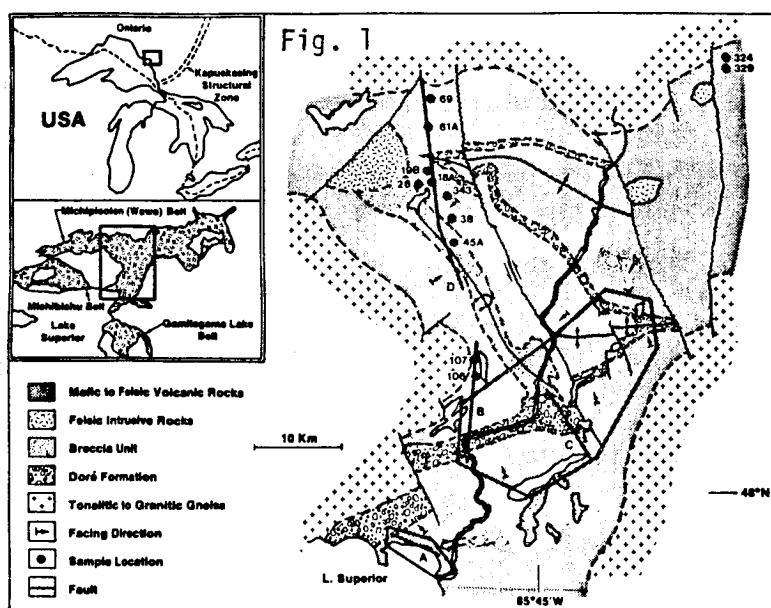
A range of depositional
environments apparently
existed for the felsic
volcanic rocks of the lower
volcanic sequence. Subaerial
non-welded massive ash flows,
shallow water accretionary
lapilli-bearing hyalotuffs
and deeper water bedded
pyroclastic deposits all have
been recognized [6,7,10].
Similarly, sedimentary rocks
that overlie the lower
volcanic sequence were
deposited in both subaerial
(braided fluvial and alluvial
fan) and subaqueous
(turbidite) environments
[11].

Fig. 2

Legend:
- MAFIC FLOWS
- FELSIC PYROCLASTICS
- FELSIC FLOWS
- IRON FORMATION
- CONGLOMERATE
- GREYWACKE, SHALE, SILTSTONE
- VOLCANIC BRECCIA
- GRANITOID

Voluminous Cenozoic rhyolitic pyroclastic deposits are erupted on continental (rather than oceanic) crust and exhibit distinctive chemical characteristics and rock associations depending on whether that crust was the site of intracontinental rifting or subduction. Three examples of Cenozoic rhyolites associated with intracontinental, extension-related tectonism are presented in Table 1. The Trans-Pecos volcanic province of west Texas represents a rift dominated by alkaline to peralkaline rocks of bimodal basalt-rhyolite composition. The rhyolites are dominated by low-silica (<75wt%) compositions that tend to be depleted in alumina and lime relative to iron and the alkalis. The Rio Grande rift of New Mexico consists of a more continuous spectrum of mafic to felsic compositions that are commonly described as calc-alkaline [14]. Rhyolitic rocks, such as the Bandelier Tuff, are dominated by high-silica compositions. The Yellowstone Plateau volcanic field represents a third extension-related rhyolite group characterized by an association with continental flood basalts and "hot spot" activity. Yellowstone rhyolites are compositionally similar to the subalkaline rhyolites of the Rio Grande rift.

Cenozoic ash flow tuffs of rhyolitic composition also are erupted in voluminous proportions in continental inner arc regions of convergent plate margins. Relative to rhyolites formed in intracontinental rifts or hot spots, inner arc subduction-related rhyolites tend to have higher ratios of alumina and lime to iron and the alkalis (> about 1.4) and a more continuous spectrum of low- to high-silica compositions. Three examples of inner arc Cenozoic rhyolites are listed in Table 2. They differ mainly with respect to whether a field association with voluminous coeval intermediate volcanics is present (San Juan field), ambiguous (Sierra Madre Occidental) or not found

TABLE 1

VOLUMINOUS CENOZOIC RHYOLITIC ASH-FLOW TUFFS

Extension-related, intracontinental suites

	Dominant SiO$_2$ range	(Al$_2$O$_3$ + CaO)/ (FeO$_t$ + alkalis)	Associated volcanics
1. Trans Pecos Volcanic Province, Texas [12]	70-75 wt.%	0.76 - 1.25	hawaiite, mugearite, trachyte
2. Bandelier Tuff, Jemez Mountains, New Mexico [13,14]	74 - 77	1.21 - 1.43	basaltic andesite to rhyodacite
3. Yellowstone Plateau Volcanic Field [15]	75 - 77	1.24 - 1.37	olivine tholeiite

TABLE 2

VOLUMINOUS CENOZOIC RHYOLITIC ASH-FLOW TUFFS

Subduction-related, continental inner arc suites

	Dominant SiO$_2$ range	(Al$_2$O$_3$ + CaO)/ (FeO$_t$ + alkalis)	Associated volcanics
1. Taupo Volcanic Zone, New Zealand [16]	69-77 wt.%	1.57 - 1.71	minor high-Al basalt to dacite
2. Mid-Tertiary Upper Volcanic Sequence, Sierra Madre Occidental, Mexico [17]	70 - 77	1.62 - 1.85	minor basaltic andesite to dacite
3. Oligocene Ash-Flows, San Juan Volcanic Field, Colorado [18]	64 - 76	1.34 - 2.05	voluminous andesite to qtz latite

(Taupo volcanic zone).

If Cenozoic rhyolites may be used as a guide, the Michipicoten lower volcanic sequence (FV1) rhyolites, which are characterized by a continuous spectrum of silica compositions and relatively high ratios of alumina and lime to iron and the alkalis (Table 3), are more

likely to be subduction-related than intracontinental rift-related. The Taupo volcanic zone and neighboring Kermadec-Tonga island arc system [19] offer perhaps the most appropriate plate tectonic analogue. At this convergent plate margin, rhyolitic pyroclastic rocks erupted from the New Zealand continental crust actually are deposited largely on the adjacent sea floor [20], which also is the depositional site for tholeiites derived from the Kermadec-Tonga island arc. The resulting ocean floor/continental slope deposits should consist of interfingering rhyolites and basalts derived independently from continental and oceanic platforms, respectively.

A similar tectonic-depositional model may explain the so-called cyclical mafic to felsic stratigraphic relationships present in the Michipicoten belt. The presence of pre-existing granitoid crust flanking the belt and the well-known compositional similarity between Cenozoic island arc tholeiites and Archean greenstone belt tholeiites [21], such as those present in the Michipicoten belt [22], support this interpretation. However, the existence of subaerial and shallow subaqueous depositional environments for some Michipicoten volcanic, volcaniclastic and sedimentary units requires either intermittent, local emergence of the volcanic pile or the existence of at least small continental blocks underlying parts of the belt.

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GEOPHYSICAL CHARACTERISTICS AND CRUSTAL STRUCTURE OF GREENSTONE
TERRANES, CANADIAN SHIELD; M.D. Thomas[1], L. Losier[2], P.C.
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A knowledge of the deep structure and geometry of greenstone belts is
fundamental to tectonic models of Archean evolution. In the Canadian
Shield long linear granite-greenstone terranes of generally low
metamorphic grade alternate with temporally-equivalent metasedimentary
belts of higher grade (Fig. 1). The focus of geophysical investigations
of these terranes has been to examine geometries and contact
relationships within individual terranes, and to look at the broader and
deeper aspects of structure and inter-terrane relationships.
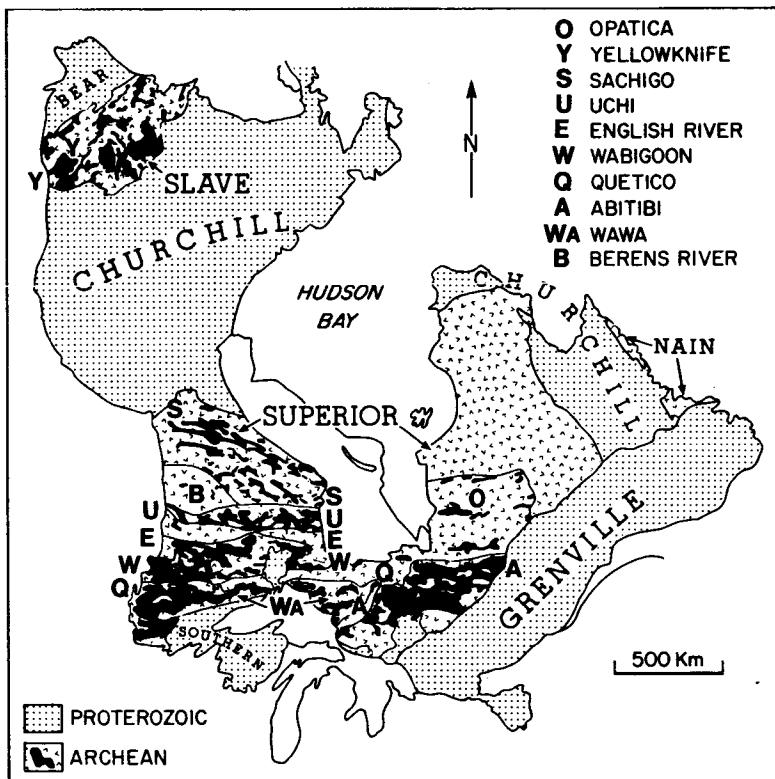

Figure 1 - Structural province map of Canadian Shield showing
distribution of greenstones, subprovinces of Superior province and
Yellowknife greenstone belt of Slave province.

Major greenstone belts are characterized by positive gravity
anomalies in the range 15-30 mGal that primarily reflect the relatively
high density mafic and ultramafic metavolcanic components (1). These
anomalies are sometimes interrupted by negative anomalies caused by
felsic plutons and are poorly developed where high metamorphic grade
basement is present and/or boundaries are gently-dipping. Modelling
reveals that many greenstone belts are more or less basin-shaped, some
having deep keels, and that their steep surface boundaries extend to
depth. Model depths of polycyclic greenstones are 2-8 km and
non-polycyclic are 3-12 km (1). The generally smaller depths of the

former have been attributed to granitic intrusion decreasing vertical extent by stoping (Fig. 2), or to listric normal faulting or thrusting (1). Models indicate abrupt changes in depth of up to ~10 km between
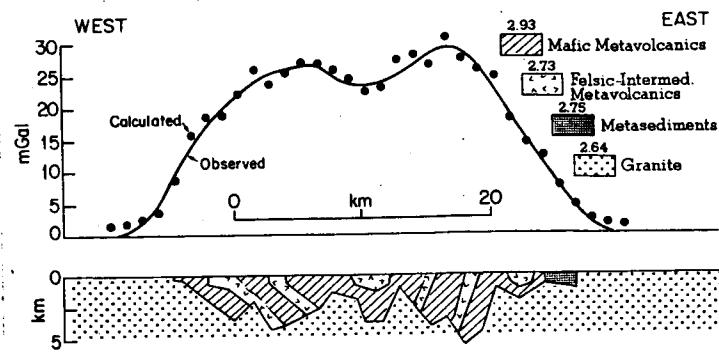

Figure 2 – Gravity model of Birch-Uchi greenstone belt, Wabigoon subprovince suggestive of removal of roots of belt by magmatic stoping and partial melting (20). Densities (g/cm^3) are indicated.

supracrustals of the Wawa greenstone and Quetico metasedimentary terranes and point to a major faulted contact (2). Granitic intrusions at and within boundaries of greenstones are associated with prominent negative gravity anomalies. Modelling indicates that they have depths ranging from 2-16 km with depths in the middle of the range being characteristic (3,4). Generally, the contacts of the granites are modelled as steeply dipping. Some granites extend several kilometres deeper than adjacent greenstones but in other cases greenstones are interpreted to underlie the granite. For example, interpretation of a combined gravity-seismic study of the Aulneau batholith of the Wabigoon subprovince suggests that it is floored by up to 10 km of greenstones (3). Gravity studies in Wabigoon subprovince have contributed to classifying granites into epizonal sheets and deep diapiric batholiths intruded in two separate periods (4).

Regionally, greenstone belts generally correspond to magnetic lows and associated granites to magnetic highs (5,6). Magnetization studies (6) indicate values that are generally < 0.05 A/m for greenstones and > 0.05 A/m for granites. Linear positive anomalies within the English River gneiss belt have drawn attention to pyroxene amphibolite gneisses, probably derived from metavolcanics (7). Their occurrence is significant in that they are in an area where volcanism is thought not to have been important. Aeromagnetic shaded relief maps have been used to assist in mapping surface geology in the Abitibi greenstone belt (8). Various features correlate with diorite-gabbro and peridotite-serpentinite intrusions, diabase dykes, major faults, iron formations and zones of contact metamorphism around granitic intrusions. The magnetic signature of the Abitibi belt, however, is not noticeably different from that of the bordering terranes. Modelling has been limited. Interpretation of a 300 km N-S profile across the Abitibi belt (8) indicates that the greenstones extend to a maximum depth of 13.6 km in the south, with an average depth of ~9 km compared to 6 km in the north (Fig. 3). This agrees with seismic refraction results that suggest the bottom of the belt dips southward increasing in depth from 6 to 14 km (9). Surface magnetic units over granites of the Wabigoon belt have been modelled as extending to the intermediate discontinuity (16-19 km) with an increase in magnetization occurring at a few kilometres depth (6). Magnetization is low or absent below the discontinuity.

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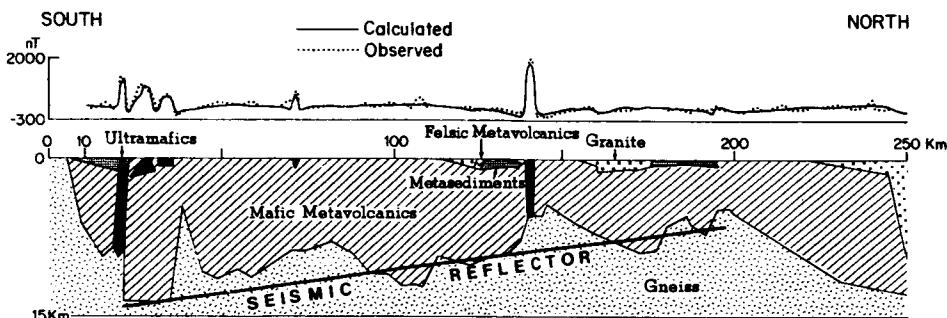
Figure 3 - Aeromagnetic (8) and seismic (9) interpretations of Abitibi greenstone belt.

Seismic reflection studies within the Aulneau batholith and adjacent greenstones (10,11) have mapped a near-vertical contact between granite and greenstone to a depth of several kilometres (confirmed by later gravity studies) and a vertical fault zone. Although there are no detectable velocity differences between the greenstones and granites, the impedance contrast is sufficient to produce recognizable reflections from the near-vertical contact. The lower surface of the batholith, as interpreted from gravity, did not produce reflections, perhaps due to its undulatory nature (12). There is also a poor correlation between the average depth of the Yellowknife greenstone belt as determined from seismic (~10 km) and gravity (~3 km) studies (13,14). In contrast, the seismic refraction survey (9) across the Abitibi belt yielded a geometry for the bottom of the belt similar to that based on magnetic interpretation (8). The seismic investigations in the vicinity of the Aulneau batholith (10,11) also detected several deep horizontal or near-horizontal reflectors. The most prominent reflectors are at intermediate depths of about 19 and 22 km and the Moho at 38 km. The three reflectors appear to be continuous beneath the granite and greenstones suggesting that complex structure, which typifies the upper crust, is absent at depth. A similar picture of the Wabigoon crust has been found by long-range refraction - wide angle reflection experiments (15,16), but in the Quetico metasedimentary belt to the south no sharp boundaries are found within or at the base of the crust which is about 40-42 km thick (16). In the English River gneiss belt to the north seismic refraction studies indicate thinner crust with an average thickness of 34 km (17). The average depth of the intermediate discontinuity remains about the same (~18 km). In detail, the Moho is upwarped by roughly 8 km in the northern part of the belt, whereas the intermediate discontinuity exhibits a complementary downwarp with an amplitude of 10 km. Re-examination of the orginal data (12) indicates that the axis of this proposed warping lies close to the northern margin of the gneiss belt where it coincides with a sedimentary basin.

Magnetotelluric investigations have been carried out in the western Wabigoon belt (18). A 3.9 km thick near-surface resistive zone under the metavolcanics is considerably less resistive (21,300 Ω-m) than one 7.4 km thick under the granitic gneiss (3,280,000 Ω-m). It suggests that crust underlying metavolcanic rocks is partially fractured and contains saline fluids and/or that the metavolcanics extend throughout the resistive zone. Heat flow studies reported from several Precambrian

shields indicate that the average heat flow in greenstones is roughly 10% lower than in crystalline terranes (19). Heat generation data from the Churchill and Superior provinces of the Canadian Shield indicate that greenstones are approximately 7 km thick.

A general conclusion is that greenstone belts are not rooted in deep crustal structures. Geophysical techniques consistently indicate that greenstones are restricted to the uppermost 10 km or so of crust and are underlain by geophysically normal crust. Gravity models suggest that granitic elements are similarly restricted, although magnetic modelling suggests possible downward extension to the intermediate discontinuity around ~18 km. Seismic evidence demonstrates that steeply-dipping structure, which can be associated with the belts in the upper crust, is not present in the lower crust. Horizontal intermediate discontinuities mapped under adjacent greenstone and granitic components are not noticeably disrupted in the boundary zone. Geophysical evidence points to the presence of discontinuities between greenstone-granite and adjacent metasedimentary terranes. Measured stratigraphic thicknesses of greenstone belts are often twice or more the vertical thicknesses determined from gravity modelling. Explanations advanced for the discrepancy include stratigraphy repeated by thrust faulting and/or listric normal faulting (1), mechanisms which are consistent with certain aspects of conceptual models of greenstone development. Where repetition is not a factor the gravity evidence points to removal of the root zones of greenstone belts. For one region, this has been attributed to magmatic stoping during resurgent caldera activity (20).

Geophysical studies in the Canadian Shield have provided some insights into the tectonic setting of greenstone belts. Much work, however, remains to be done, particularly in the use of geophysics in evolutionary models of greenstone development. Future needs include detailed, integrated studies, the introduction of relatively new methods such as Vibroseis seismic reflection, greater use of magnetotellurics and the application of other electromagnetic methods such as very low frequency (VLF) surveys.

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VOLCANOLOGICAL CONSTRAINTS ON ARCHEAN TECTONICS P. C. Thurston and
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Terranes within Archean shields can be classified as granite-greenstone
megabelts, contemporaneous sedimentary megabelts (1) and basement enclaves
within either of the above (2,3). Stratigraphic and geochronological work in
the Superior Province has shown the granite-greenstone megabelts represent
proximal volcanism, sparse deep water clastic sedimentation, and late alluvial
fan-submarine fan sedimentation(4). The sedimentary megabelts represent
stratigraphically equivalent deep water sedimentation of wacke-pelite couplets
(5), submarine fan conglomerates and minor distal facies volcanism(6). Basement
enclaves include meta-igneous and metasedimentary gneiss and fragmented
metavolcanic relics with poorly preserved primary textures.(2,7).
Greenstone belts of Australia have been subdivided into >3 Ga platform-phase
greenstones and <3Ga rift-phase greenstones (8). Platform phase units are basal
komatiite flows and tholeiitic flows with an upper unit of minor pyroclastics.
Volcanism in the platform phase is typified by abundant pillowed amygdular
flows, overlain by minor airfall tuff and relatively distal debris flow
volcaniclastic units. Sedimentary units include chert, quartzite, and
stromatolitic carbonates with minor wackes, indicative of shallow water
platform sedimentation (9). Examples in the Superior Province, generally about
3 Ga old (10) include quartz-rich wackes in the lower sequence at North Spirit
Lake in the Sachigo Subprovince (11), quartzites with fuchsite clasts in the
lower sequence of the Wabigoon Subprovince at Armit Lake (12), and carbonate-
rich sediments in the Lumby Lake greenstone belt (13) within the Wabigoon
Diapiric Axis basement enclave (3). Volcanologically one can conclude from the
thickness of the shallow water volcanic rocks and sediments that accumulation
took place on a shallow platform (9) and as well, large scale subsidence kept
pace with the rate of accumulation of volcanic rocks.
Rift-phase greenstones are relatively deep water amygdule-poor pillowed
tholeiites succeeded upward by vesiculated pillowed flows and calc-alkaline
pyroclastic and volcanoclastic units (8). Considering the maximum water depth
for pyroclastic erutions (14) and the thickness of pyroclastic sections in many
rift-phase greenstone belts, Ayres (15) has suggested many Plinian eruption
columns became subaerial. Classically (15) most Archean pyroclastic units were
considered to have been deposited subaqueously. Recent studies have shown
however that many Archean pyroclastic units were deposited subaerially (16,15).
Sedimentologic studies of rift-phase greenstones show some deep-water clastic
deposits(17), but increasingly shallower water deposits (alluvial fan) at
stratigraphically high levels. The structural pattern in rift-phase belts is
alternating synclinoria and anticlinoria either breached by diapirism or
sheared out (8). Most Superior Province greenstone belts younger than 2.9-2.8
Ga (18) are probably rift-phase based on the following. a)structural style with
synclinoria dominating with only rare dome and basin structures. b)
Quartz-rich and carbonate rich sedimentation is scarce in the Abitibi (4),
kWabigoon(20), and younger (<2.9Ga) sequences of the Uchi (18) and Sachigo
(11,18) Subprovinces. c) Volcanism is typified by bimodal tholeiite-rhyolite
sequences (21) with shoaling upward attributes (18,4). Evidence for small
scale operation of rift-related volcanism is seen in the Six Mile Lake cycle at
Sturgeon Lake (22) where a tholeiitic basalt-calcalkaline rhyolite sequence
2755 Ma (23) is rich in incompatible elements relative to later sequences and
is cut by abundant mafic dikes assumed to feed the uounger (2718Ma) (24) cycles
related to wide-scale rifting.
A survey of volcanic cyclicity (20) reveals the following types of cycles
within the Superior Province. (+=fractionation relation; -=no fractionation
KOM =komatiitic; TH=tholeiitic; CA=calc-alkaline; ALK=alkaline magma clans

1)KOM Perid Kom +dacite   4)TH bas + andes - Ca bas + rhy - ALK

2)KOM Perid.kom - TH bas + rhy -   5)CA bas + rhy
   -CA bas + rhy -ALK

3)TH bas + andes - Th andes - Ca dac 6)TH bas - CA dac + rhy - THbas
  +rhy
Increasing stratigraphic height to the right in each entry.
Cycle types 3,4, and 5 above were formerly thought to represent
fractionation sequences,but recent work has shown that many are bimodal(21).
The fact that the above cycle types are bimodal has profound volcanologic and
petrogenetic implications in that the bimodalism is not simply the paucity of
intermediate composition magmatic liquids. Trace element geochemistry and
field evidence suggests, when corrected for unerupted volume in zoned magma

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chambers,and loss of vitric fines in high level winds during Plinian eruptions
are made, preserved volumes of felsic volcanics in the Archean represent $\pm$ 15%
of the original felsic magma (21). In effect, we concluded that Archean
bimodal volcanism represents subequal volumes of mafic and felsic magma which
are involved in greenstone belt volcanism.

Determination of paleoenvironment (above), eruption type, eruption rate,
magma chamber size and type, developmental processes, and the life span of
individual volcanoes places many genetic constraints on greenstone belt
tectonics. In mafic sequences subequal volumes of pillowed and massive flows
(18) suggest eruption by sheet flow processes (25) dominate over eruption from
shield volcanoes (18). In felsic sequences the volumetric dominance of
ignimbrites (21) and the notion that sedimentary basins contain large amounts
of tephra suggest Plinian eruptions were dominant in the Archean. Many Plinian
eruptions produced subaerial deposits on local volcanic islands (18.19,15).
Vulcanian eruptions are subordinate, they produce less widespread deposits -
examples include the Skead Group (26) and the Lake of the Woods area (27).
This eruption type is often the result of less volatile-rich magmas relative to
Plinian systems (28) interacting with near-surface water. The deposits are
generally less widespread in extent than many Plinian deposits.

Eruption rates of Archean volcanoes can be determined in an approximate
and indirect fashion. Sheet flows (25) a greater mean flow thickness than in
Phanerozoic analogues (18) and the presence of lava plains (29) in Archean
mafic sequences suggest a more rapid eruption rate than in Phanerozoic
analogues (30). Phanerozoic ignimbrite systems have volumes in the 10^1-10^2 km^3
range (31) with exceptional examples in the 10^3-10^4 km^3 range (31,32).
Phanerozoic felsic volcanoes had a life-span generally not exceeding 1.5 Ma
(18) but many Archean felsic edifices apparently existed for 10-20 Ma(18).

The preserved volume of felsic ignimbrites (recalculated to compensate for
unerupted material and loss of vitric fines, but ignoring compaction)
suggests existence of felsic magma chambers on the order of 10^3 km^3 (21)
rivalling those of the largest Phanerozoic systems (28,29). When integrated
with data on the lifespan of Archean volcanoes of 10-20 Ma, Archean felsic
eruption rates were large, but not as large as those seen in Archean mafic
systems.

Volcanological and trace element geochemical data can be integrated to place
some constraints upon the size, character and evolutionary history of Archean
volcanic plumbing, and hence indirectly, Archean tectonics. The earliest
volcanism in any greenstone belt is almost universally tholeitic basalt.
Archean mafic magma chambers were usually the site of low pressure
fractionation of olivine, plagioclase and later Cpx+ an oxide phase during
evolution of tholeitic liquids (33 and references therein). Several models
suggest basalt becoming more contaminated by sial with time (33,34). Data in
the Uchi Subprovince shows early felsic volcanics to have fractionated REE
patterns (33)followed by flat REE pattern rhyolites. This is interpreted as
initial felsic liquids produced by melting of a garnetiferous mafic source
followed by large scale melting of LIL-rich sial (33). Rare andesites in the
Uchi Subprovince are produced by basalt fractionation, direct mantle melts and
mixing of basaltic and tonalitic liquids(33). Composite dikes in the Abitibi
Subprovince (35) have a basaltic edge with a chill margin, a rhyolitic interior
with no basalt-rhyolite chill margin and partially melted sialic inclusions.
Ignimbrites in the Uchi (16) and Abitibi (36) Subprovinces have mafic pumice
toward the top. Integration of these data suggest initial mantle-derived
basaltic liquids pond in a sialic crust, fractionate and melt sial. The
inirial melts low in heavy REE are melts of mafic material, subsequently
melting of adjacent sial produces a chamber with a felsic upper part underlain
by mafic magma.
Compositional zonation of the overlying felsic magma develops with time (31),
resulting in Plinian eruption through roll over (37) or volatile
supersaturation(38).

Numerous arguments suggest widespread volcanism-related subsidence kept pace
with the rate of eruption: a) The preservation of felsic sequences rather than
the rapid erosion common in Phanerozoic terranes (39) b)Minimum water depth for
pyroclastic activity (14) vs preserved stratigraphic thickness of subaqueous
pyroclastic units (15)i.e. sections are much thicker than maximum water depth

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for eruption - therefore subsidence occurred. c) Lateral extent of 30-50 km for stromatolitic carbonates (40) in the Uchi subprovince,lateral extent of 30-50 km for shallow water silicified evaporites (41) and lateral extent and high eruption rate for shallow water environment mafic plains would have rapidly become subaerial unless subsidence kept pace(18). Isostatic calculations (42,43) suggest lava plain eruptions produce lesser crustal loading than central vent eruptions and less isostatic subsidence. Models involving sialic substrate to lava plain systems produce (42) sufficient subsidence to just maintain volcanic piles at sea level. Therefore we conclude a) subsidence kept pace with volcanism, b)subsidence was regional in extent, c) it is difficult to envision a sagduction style of subsidence (44) producing subsidence over a large area consistent with the great areal extent of the main contributor to the subsidence- the mafic lava plains. Subsidence was more rapid during mafic volcanism slowing during felsic volcanism.

The great volumes of Archean rhyolites and bimodal nature of rift-phase volcanism mitigates against an island arc or back-arc basin analogue where rhyolite is scarce (39 and references therin). Both continental arcs and continental rifts have sufficient volumes of felsic volcanism to compare to greenstone belts. The sediment-filled grabens associated with the Rio Grande Rift (45) offer a possible modern analogue as do the continental intra-arc depressions (39).

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THE DEHYDRATION, REHYDRATION AND TECTONIC SETTING OF GREENSTONE
BELTS IN A PORTION OF THE NORTHERN KAAPVAAL CRATON, SOUTHERN AFRICA; D.D. van
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I. GEOLOGICAL CHARACTERISTICS OF ARCHAEAN CRUST

Two types of Archaean crust are commonly recognized: 1) low-grade granite-
greenstone terranes and 2) high-grade gneiss terranes. Generally high-grade
terranes are viewed as being distinct from typical low-grade granite-greenstone
terranes with which they are often associated. Three models have been proposed
to explain the relationship between the two types of terranes (1). The first
model considers high-grade terranes to be basement to younger low-grade green-
stone belts. The second model regards the evolution of low- and high-grade
terranes as coeval but in different environments. The third model (as sup-
ported here) is that the low-grade granite-greenstone terranes and the high-
grade gneiss terranes represent cross sections through Archaean crust which
was subjected to plate tectonic processes and, in particular, to the collision
of granitoid continents. The detailed examination of well-exposed Archaean
terranes at different metamorphic grades, therefore, is not only an important
source of information about the crustal levels exposed, but also is critical
to the understanding of the possible tectonic and metamorphic evolution of
greenstone belts with time. Integration of this information and disciplined
acquisition of critical data from suitable areas will provide the necessary
answers to the applicability of plate tectonics in these times.

II. RELATIONSHIP BETWEEN ARCHAEAN LOW- AND HIGH-GRADE TERRANES IN THE NORTHERN KAAPVAAL CRATON

Many features of a metamorphic and deformational transition from a typical
low-grade granite-greenstone terrane to a high-grade gneiss terrane are illus-
trated in the crustal section of the northern portion of the Kaapvaal Craton
over the 60 km between the Pietersburg Greenstone Belt and the granulite facies
Southern Marginal Zone of the Limpopo Belt (Fig. 1). In this section, steeply
dipping, typical greenstone belt lithologies occur at higher and higher grade
moving from south to north. In the south, the Pietersburg Belt comprises an at
least 3450 Ma mafic, felsic and ultramafic volcanic and volcano-sedimentary as-
semblage (the Pietersburg Group) unconformably overlain by a sedimentary assem-
blage (the Uitkyk Formation), probably deposited between about 2800 Ma and 2650
Ma. The Pietersburg Group is surrounded by the approximately 3500 Ma tonalitic
and trondhjemitic Baviaanskloof Gneiss and is intruded by the approximately
2800 Ma Hout River Gneiss. All these units are intruded by approximately 2650
Ma, largely undeformed, granodioritic plutons. Metamorphic grade within the
Pietersburg Belt increases from greenschist facies in the southwestern and cen-
tral parts to amphibolite facies in the northeast, consistent with the regional
metamorphic pattern.

North and northeast of the Pietersburg Belt are situated mafic, felsic and
ultramafic volcanic and sedimentary rocks of the Rhenosterkoppies and Suther-
land Greenstone Belts. Both belts are surrounded by the Baviaanskloof Gneiss.
The ages of the lithologies within these belts are unknown but both belts have
been metamorphosed under amphibolite facies conditions.

Typical greenstone belt lithologies can be followed uninterruptedly across
the transition from amphibolite facies to granulite facies within the Southern
Marginal Zone where they are highly attenuated and boudinaged in the Baviaans-

kloof Gneiss. These assemblages are intruded by the approximately 2650 Ma deformed Matok granodioritic pluton and the undeformed 2450 Ma Palmietfontein granite.

The transition from the low-grade granite-greenstone terrane to the Southern Marginal Zone is not only reflected by an increase in the grade of metamorphism but also by an increase in the intensity of deformation. The structural grain of the entire region trends east-northeast with an almost vertically dipping schistoscity or gneissoscity. In the Southern Marginal Zone, the distended nature of the granulitic greenstone remnants is in sharp contrast to the more continuous outcrop pattern of the greenstone lithologies to the south.

III. METAMORPHIC EVOLUTION OF THE SOUTHERN MARGINAL ZONE

The two-fold division of greenstone belts proposed by Binns et al. (2) for the eastern Yilgarn Block can be applied to the northern portion of the Kaapvaal Craton. Areas of static metamorphism are of low-grade with little internal strain and are restricted to the central and southwestern portion of the Pietersburg Belt. Areas of dynamic metamorphism are of medium- to high-grade and are highly strained. These areas include the northeastern part of the Pietersburg Belt as well as both the Rhenosterkoppies and Sutherland Belts. Areas with intermediate characteristics occur within both the Pietersburg and Sutherland Belts. Geophysical data indicate that the thickest greenstone successions occur under areas of static metamorphism.

Peridotitic komatiite occurring in areas of static metamorphism within the Pietersburg Belt is characterized by olivine phenocrysts which have been completely replaced by serpentine or chlorite. In equivalent rocks subjected to dynamic metamorphism, olivine phenocrysts and spinifex textured olivine remain unaltered. This relationship implies that the volcanic rocks in the high-grade domains did not suffer alteration equivalent to that in the low-grade domains. Igneous olivine, therefore, transformed directly to metamorphic olivine without undergoing prior serpentinization. Regional metamorphism in the northern Kaapvaal Craton was not progressive but rather the main phase of recrystallization did not occur until peak metamorphic conditions had been established within formerly little altered greenstone sequences (2).

The crustal behavior of the entire northern portion of the Kaapvaal Craton must have been consistent with the observation that the rocks of the Southern Marginal Zone were depressed into deep crustal levels. This movement implies that the low-grade terranes were probably depressed in a sympathetic manner. In the Southern Marginal Zone (Fig. 2), the maximum prograde conditions (P>9.5 kb and T>800°C) reached during this tectonic event are recorded by the assemblage garnet + hypersthene + quartz + plagioclase +/- kyanite +/- biotite in metapelite. These conditions were followed by rapid, nearly isothermal, decompression between approximately 2700 Ma and 2650 Ma, recorded by decompression textures of cordierite and hypersthene after garnet. P-T conditions of this decompression event were T=800°C and P decreasing to 7.0 kb. The Matok pluton was emplaced during the isothermal decompression. The southern margin of this dehydrated terrane was then subjected to a regional encroachment of CO_2-rich hydrating fluids before approximately 2450 Ma, the time of emplacement of the Palmietfontein granite. This encroachment produced the retrograde orthoamphibole isograd defined by the reactions: hypersthene + quartz + H_2O = anthophyllite and cordierite + H_2O = gedrite + kyanite + quartz. These reactions occurred at T=650°C to 600°C and a total P less than 6 kb at $PH_2O = 0.2P_{total}$. Completely hydrated and recrystallized rocks south of this isograd are characterized by the assemblage anthophyllite + gedrite + kyanite + biotite + quartz + plagioclase. The fluids responsible for rehydration are believed to have been derived from hydrated granite-greenstone lithologies.

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A parallel P-T-time scenario affected the rocks of the Central Zone of the Limpopo Belt immediately to the north, indicating that the present erosional level of both the Central and the Southern Marginal Zones is an isofacial surface of constant P-T-time conditions.

IV.   TECTONIC DEVELOPMENT OF THE NORTHERN KAAPVAAL CRATON

The lack of correlation between metamorphic grade and the distribution of granitoid plutons precludes the development of the regional metamorphic pattern as a result of granite emplacement.   Also, the overall increase in the grade of metamorphism and the distribution of relict igneous minerals suggest that the metamorphic evolution of the high-grade Southern Marginal and Central Zones of the Limpopo Belt were coeval with that of the adjacent granite-greenstone terrane to the south.   The ubiquity of 2700-2600 Ma ages throughout the northern and central portions of the Kaapvaal Craton suggests that the metamorphic pattern arose during a very wide-spread tectonic event which was linked with the generation and remobilization of granitic rocks.   This relationship precludes the existence of different geothermal gradients for greenstone belts and for high-grade gneiss terranes and suggests that the northern portion of the Kaapvaal Craton represents a cross section through Archaean crust.

Gravity and resistivity data indicate that the Pietersburg, Rhenosterkoppies and Sutherland Greenstone Belts are shallow features, rarely exceeding 5 km in depth.   5 km depth is in marked contrast to the great thicknesses of various steeply dipping lithologic successions in the area measured across the stratification (up to 25 km) and to the depth at which these rocks were metamorphosed.   These observations indicate that major crustal thickening took place.   The existence of high-grade assemblages at the surface overlying crust of thickness of 40 km indicates that thickened crust in excess of 80 km existed approximately 2650 Ma ago.   It is proposed that this crustal thickening was achieved by thrusting in a zone of crustal convergence in which two or more continental fragments collided, analogous to the tectonic activity presently going on in the Himalayas.   This model is supported by isotopic and lithologic evidence for the existence of exotic terranes in this area.   Possibly some of the thrust faults along which this thickening occurred may be recognized in modified form as "straightening zones".   The depressed portions of the crust were highly deformed during this compressional period and the rocks were subjected to high-grade metamorphism so that new isograds were established, discordant to the imbricate thrust structures.   Originally lower crustal dehydrated rocks were largely unaltered metamorphically at great depth but upper crustal hydrated rocks that were buried to similar depths underwent rapid metamorphism.   Large volumes of granitoid rocks were also created.   Later uplift giving rise to the observed decompression textures (+/- 0.5 cm per year) was achieved by a combination of melt enhanced "surge tectonics" (3), isostatic adjustment of the thickened crust to erosion and the collapse of the thickened crust under its own weight.   The expression of this uprising and deforming mass is thrusting radially out of the high-grade zone recognized today as the Limpopo Belt.   Continental collision and mountain building about 2650 Ma ago must be a significant factor in the formation of the unparalleled gold mineralization of the Witwatersrand Basin immediately to the south.

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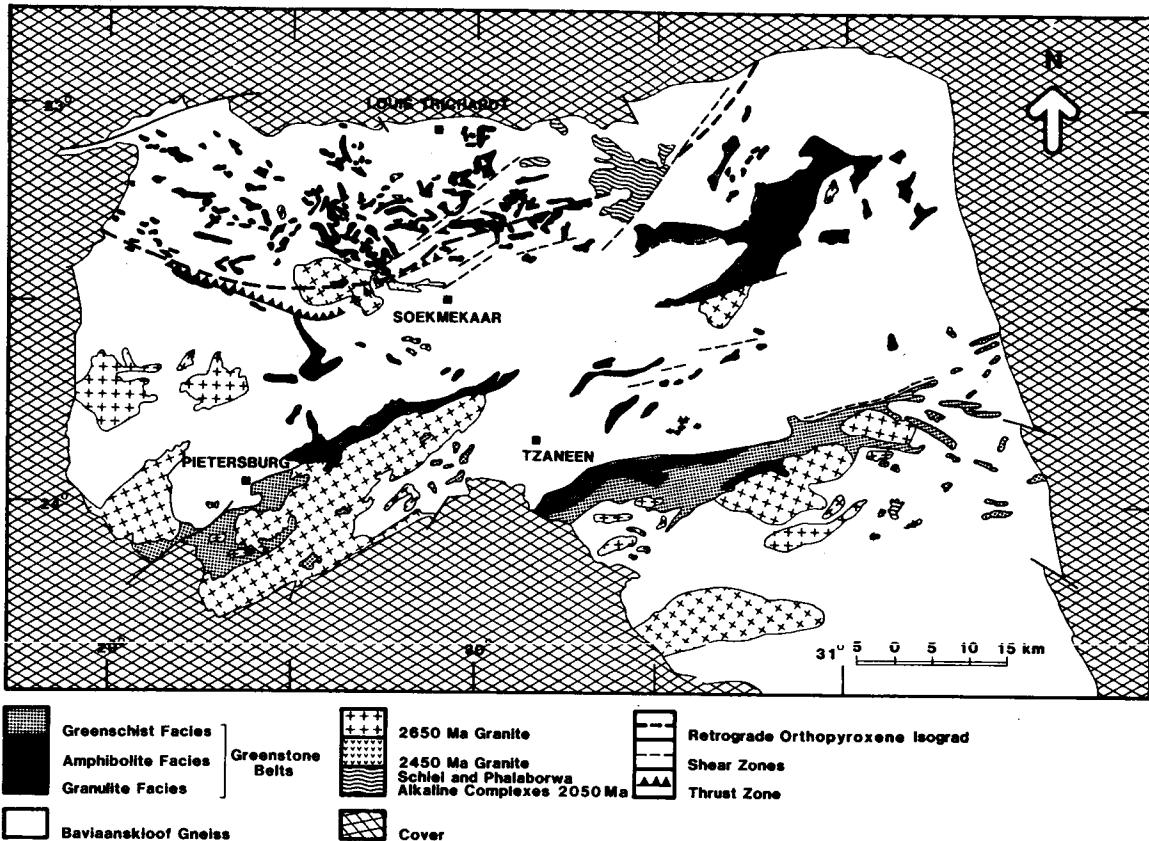
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Fig. 1: Generalized metamorphic map of the northern portion of the Kaap-
vaal Craton including the Southern Marginal Zone of the Limpopo
Belt.

Fig. 2
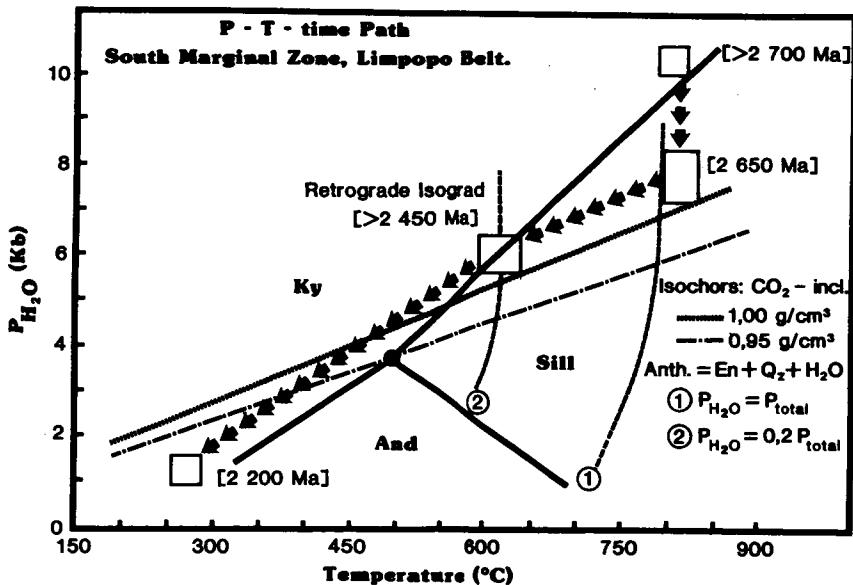
GREENSTONE BELTS: THEIR COMPONENTS AND STRUCTURE.

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Although in common geological usage there is considerable ambiguity over the definition of greenstone belts which are historically regarded as long and narrow in shape, Archaean in age and composed of volcanic and sedimentary sequences at greenschist facies. This definition remains true for many of what are commonly regarded as greenstone belts but others differ significantly, particularly in shape and metamorphic facies. For this reason the term 'succession' is prefered for greenstones which are not particularly linear. In the following discussion it is our intention to maintain 'greenstone' as a useful term and for that reason we specifically aim to exclude high-grade supracrustal gneiss terrains such as those of the central zone of the Limpopo belt and early Precambrian supracrustal sequences such as the 3Ga Pongola, the 2.7Ga Witwatersrand and the 2.4Ga Ventersdorp from any definition of greenstone successions. We also aim to include all commonly accepted greenstone successions. The following points are of relevance to the definition of greenstone belts:

1. Most commonly accepted greenstone successions are of Archaean age but a few younger belts have been reported from Wisconsin, USA (1) and northern Quebec, Canada (2).
2. Although many greenstone successions are long, linear and narrow (e.g. Pietersburg and Murchison, Kaapvaal craton) many others have more irregular shapes (eg. Bulawayan, Zimbabwe craton and Pilbara, Western Australia). The word 'belt' therefore is inappropriate for some greenstone successions.
3. Volcanic rocks are ubiquitous components whereas sediments may be of secondary importance. The volcanics frequently include komatiitic rocks. Intrusive igneous rock units such as layered complexes, dykes and sills may be present.
4. Greenstone successions occur at metamorphic conditions from sub-greenschist to granulite facies and the colour prefix, refering to the greenschist facies, is unfortunate.
5. Deformation intensity within the greenstone successions is variable.
6. Greenstone successions are always intimately associated with and surrounded by trondhjemite-tonalite-granodiorite-granite granitoids.

We tentatively suggest the following definition:
Greenstone successions are the non-granitoid component of granitoid-greenstone terrains. Volcanic rocks are an essential component, some of which are usually komatiitic. Sedimentary rocks are commonly present and igneous intrusive units may exist. The greenstone successions are linear to irregular in shape and where linear they are termed belts. The greenstone successions may occur at all metamorphic facies and are heterogenously

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deformed. Most greenstone successions are Archaean in age.

Greenstone successions comprise a wide variety of rocks, dominated by volcanics, which are usually altered and deformed. Alteration of volcanic and other rock types is manifested by hydration with variable silicification (3), carbonate-isation (4) or silica loss (5) as well as isochemical metamorphism. Alteration itself is temporally and spatially variable, Smith and Erlank (6) have described possible early sea floor alteration of komatiitic rocks from Barberton and carbonate-isation in Murchison is patchy and syn- to post-tectonic. This alteration constrains identification of original rock-types and the use of whole rock chemistry. This restriction added to the problems of equating area of surface outcrop with rock volume means that estimates of greenstone lithological proportions must be treated circumspectly. However, greenstone successions commonly comprise the following primary lithologies: komatiitic, mafic and felsic volcanics, cherts, banded iron formations, shales, graywackes and quartz arenites. Less commonly limestones (including stromatolites), arkose, ultramafic and mafic layered complexes, quartz-feldspar porphyries and quartz tholeiite dykes are present.

The identification of the environment of emplacement of greenstone igneous rocks is highly problematic. Subvolcanic intrusions exhibit many features almost indistinguishable from true lavas. Skeletal crystal growths, commonly grouped under the all-embracing term of 'spinifex', are an important textural form in these rocks and these textures, in abundance, are restricted to Archaean greenstone successions. These textures are indicative of rapid crystal growth under supersaturated conditions (7) and need not be restricted to lava flows. In fact, the inordinatly thick cumulate zones associated with some spinifex-bearing rock-types preclude these being lava flows in the currently accepted sense and the non-genetic term 'cooling unit' has been used to describe these layered rocks which may represent lava flows or subvolcanic intrusions. The recognition of crescumulate type crystal growth and rhythmically developed spinifex units indicate a variety and complexity of mechanisms which have given rise to these textures and criteria should be established to permit the environment of emplacement to be determined more precisely. Symmetry of structures and spinifex textures encountered in some units may be indicative of dyke emplacement.

Until recently, greenstone research was largely oriented towards deducing a unifying model, subsequently heterogeneity has become the key-word. In essence, greenstone belts are of different ages and formed in different tectonic situations. Groves and Batt (8) recognise both younger and older greenstone successions in Western Australia in two distinct environments, determined on the basis of volcanic constituents, sedimentary facies, mineral deposits and tectonic style, to which they gave a

genetic interpretation as rift-phase or platform-phase greenstones. Whereas this is a major development in understanding Australian greenstones the division of other greenstone successions into rift- and platform-phase is tenuous, particularly for those of the Kaapvaal craton. The Murchison greenstone belt, for instance, has characteristics of both rift- and platform-phase greenstones.

The Barberton greenstone belt, comprising the lower komatiitic to felsic units of the Onverwacht Group and overlying deep water sediments of the Fig Tree Group, probably represents a rift-phase (8) and the overlying Moodies Group with shallow water quartzites and banded iron formation is typical of a platform-phase greenstone belt. However, herewithin lies an important observation on greenstone successions: the environment of formation can vary within a greenstone. This variation may be due to either:
1. A progressive evolution in environment. Eriksson (9) has described the Fig Tree to Moodies group evolution of the Barberton greenstone belt in terms of an evolving back-arc, or passive continental margin.
2. The superposition of different environments which are temporally separate and manifested in the field by an unconformity.
or 3. Some or all of the units are allochthonous and represent spatially and/or temporally diverse environments now tectonically juxtaposed.

Another aspect of the heterogeneity is the recognition of both continental and oceanic environments. The Mberengwa (Belingwe) greenstone belt of Zimbabwe rests unconformably on granitic rocks (10, 11, 12). Basement has also been inferred to exist beneath other greenstone belts in Australia, Canada and India (13, 14, 15). Major layered igneous complexes such as Dore Lake (16) and the Rooiwater, Murchison greenstone belt (17), are a significant component of some greenstone belts. These complexes have minor ultramafic components, anorthosite-gabbro layers, magnetitite layers and a highly differentiated and sodic granite. These complexes are analogous to bodies such as the Bushveld and are intrusions in a continental environment.

In contrast to the continental environment of some greenstone successions no proven continental basement exists at the base of the Barberton greenstone belt and the Onverwacht Group may be partially of oceanic origin (18). In addition, some ultramafic complexes may also be ophiolitic (19). De Wit and Stern (20) have recognised a possible sheeted-dyke complex in the Onverwacht group. Support for the obducted oceanic origin for some greenstone rocks comes from the recognition of podiform alpine-type chromites at Shurugwi (Zimbabwe) (21, 22) and at Lemoenfontein (Kaapvaal craton) (23). These have textural and chemical characteristics similar to those recognised in

ophiolitic complexes of Phanerozoic age.

Historically greenstone structures were regarded as simple synformal belts between sub-circular rimming granitoid domes. This relationship has given rise to genetic interpretations that greenstone belts are pinched-in synformal keels between domal or diapiric granitoids or between granitoid domes which are the result of interference folding (24). Unfortunately the paucity of detailed structural observations and accurately determined stratigraphic successions mean that few of the assumed synforms are proven.

In the Kaapvaal craton the Murchison, Pietersburg, Sutherland, Rhenosterkoppies, Amalia and Muldersdrift belts lack a gross synformal structure. At Barberton the greenstone succession comprises several synformal structures separated by steep reverse faults (25). De Wit (26) and Lamb (27) have recently described thrusts, some of which emplace Onverwacht volcanics over Moodies sediments. The suggestion of Anhaeusser (28) that deformation structures within the Barberton greenstone belt can mostly be related to granitic diapirism is at variance with the observed thrust structures and evidence presented by Ramsay (25), Roering (29) and Burke et al. (30) who note deformation structures prior to granite intrusion, intrusive granite contacts oblique to deformation structures and an absence of deformation structures within the greenstone directly related to those in the surrounding granitoids.

We suggest that whereas broadly synformal belts may exist this is not a characteristic of greenstone belts. Many of the intrusive granitoids are undoubtedly domal but intervening greenstone belts are not necessarily synformal and the role of diapirism in controlling the structure of greenstone successions may be over-emphasised.

In deducing the overall large-scale structural characteristics of greenstone successions the following general observations may be relevant:
1. Contacts with the surrounding granitoids can be either tectonic (31) or intrusive with dykes and veins of granitic rock in the greenstone belts and a static high T/low P metamorphism near the greenstone contact with the granitoids suggesting contact metamorphism by igneous intrusion.
2. Geophysical evidence from a number of belts suggests they are shallow with vertical depth extents rarely more than 10km and usually less than 5km (32, 33), figures considerably less than the proposed stratigraphic thicknesses of these belts. This shallow depth extent suggests no simple rotation of the usually upright greenstone belt but instead a truncation which may be a major decollement zone, recumbent syntectonic granite or a late intrusive contact.
3. Recumbent fold structures and possible thrusts are relatively

common and have been described from greenstone successions of the Zimbabwe craton (34, 35), of the Kaapvaal craton (25, 26, 27), of the Western Australia shield (36, 37) and the North American shield (38).
4. Greenstone successions occur as either linear belts or as irregular shaped units comprising arcuate arms.
5. Late-deformation structures and the present disposition of primary layering structures in the greenstone successions are usually upright.

Greenstone successions are composed of deformed and metamorphosed (including metasomatised) rocks. However despite the obvious difficulties, many authors have proposed stratigraphies for greenstone belts, but some have deduced total stratigraphic thicknesses dramatically in excess of those predicted by currently accepted models for basin formation (39, 40). Greenstone successions such as Barberton with 17 to 23km (41), Pietersburg with 21.4km (41) and Abitibi with over 30km (42) or up to 45km (43) total stratigraphic thickness contrast with both thinner sequences from other greenstone and non-greenstone early Precambrian supracrustal sequences such as the Witwatersrand. It is the greenstone successions with large stratigraphic thicknesses which are invariably at sub-greenschist or greenschist facies and without the high grades of metamorphism that would be expected at the base of these sequences. These thicknesses represent one of the challenging problems in greenstone geology.

Possible explanations for the large stratigraphic thicknesses are as follows:
1. They are an artifact of combining separate sections into a composite section or are oblique sections.
2. That incorporated within the greenstone belt and incorrectly interpreted as part of the stratigraphy are layered igneous complexes, sills and tectonically rotated dykes.
3. The stratigraphic sequences are in fact related to two or more spatially superimposed but temporally separate and essentially unrelated events. In the Barberton greenstone belt granite cobbles in a Moodies Group conglomerate have yielded zircons giving ages of 3.15Ga (44) contrasting with ages of 3.54Ga (45) for the stratigraphically lower Onverwacht volcanic rocks. A major phase of granite emplacement separates these two dates and a major unconformity may exist at the base of the Moodies Group.
4. They are not true stratigraphic sections but are structurally repeated by imbricate thrusting and/or folding. To achieve significant structural repetition by thrusting, folding or both requires major recumbent tectonics on or above a decollement plane.

Whilst explaining large stratigraphic repetition the recumbent thrust-fold model also predicts metamorphic conditions at the base of the pile initially at high P/low T and with

thermal relaxation to medium pressure facies. Bickle et al. (46)
have reported such rocks from the Yilgarn and similar staurolite-
kyanite-bearing rocks occur in the Murchison greenstone belt.
However the very large apparent stratigraphic thicknesses with
associated sub-greenschist or greenschist metamorphism remain
unexplained by horizontal thrust-nappe tectonics. These may
however be explained by repetition above a flat decollement in an
imbricate stack with associated folding. In this situation the
stratigraphy is turned on end and multiply repeated but the
structure remains shallow. Zones of cyclic repetition should be
investigated to determine if the cyclicity is real or the result
of imbricate stacking. Examples of this type of structural
stacking resulting in repetition are provided by Coward et al.
(35) from Matsitama, Zimbabwean craton, Botswana and Martyn (37)
from the Kalgoorlie area in the Norseman-Wiluna greenstone belt
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THE STRATIGRAPHY OF THE STEEP ROCK GROUP, N.W. ONTARIO, WITH EVIDENCE OF A MAJOR UNCONFORMITY

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The Steep Rock Group is exposed 6km north of Atikokan, 200km west of Thunder Bay. It is situated on the southern margin of the Wabigoon Belt of the Archaean Superior Province, N.W. Ontario. Reinvestigation of the geology of the Group has shown that the Group lies unconformably on the Marmion Complex to the east.

This unconformity has been previously suspected, from regional and mine mapping but no conclusive outcrop evidence for its existence has as yet been published.

The strike of the Group, comprised of five formations, Basal Conglomerate, Carbonate, Ore Zone, Ashrock and Metavolcanics is generally north-northwest dipping steeply to the southwest. Of the 7 contacts between the Steep Rock Group and the Marmion Complex, 3 expose the unconformity (the Headland, S. Roberts Pit, Trueman Point), and 4 are faulted.

At the Headland poorly sorted metaconglomerate with angular clasts of quartz, tonalite and fine-grained mafic material (dykes and remnant xenoliths) overlies mafic tonalite, with no evidence for a fault or an intrusive contact.

At the S. Roberts Pit, poorly sorted metasandstones dip steeply to the west overlying pale greenish-white weathered mafic tonalite. The metasandstones pass upwards within 20cm to massive dark grey carbonate.

At Trueman Point, in an exposure similar to the S. Roberts Pit, coarse angular metasandstone overlies tonalite. However, the contact here is more diffuse with the top metre of the tonalite breaking down to form a regolith of angular quartz grains (1-4mm) in a sericite matrix. This matrix is similar to the matrix in the overlying metasandstone (Fig. 1). These three outcrops demonstrate unequivocally that the Steep Rock Group was laid down on the underlying Marmion Complex, which is circa 3 Ga old (Davis et al, 1986).
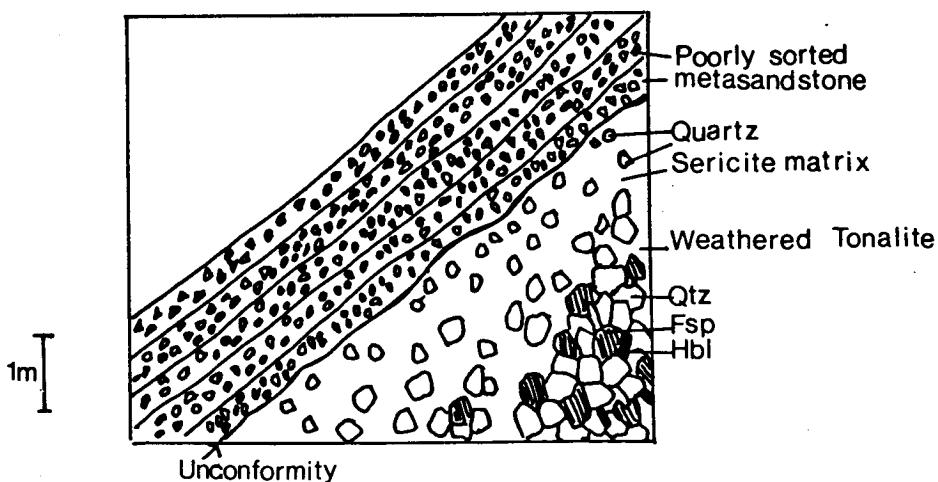

Figure 1: Drawing to illustrate the unconformable contact at Trueman Point

   Overlying the Basal Conglomerate (0-150m) is the Carbonate Formation
(0-500m) throughout which stromatolites extensively occur.
   The Carbonate is a laminated dark bluish-grey massive rock with major
zones of breccia developed close to fault zones and dykes, which are
thought to be feeders for the overlying volcanics.
   From a study of 11 stromatolitic outcrops (Wilks and Nisbet, 1985), a
crude stratigraphy within the Carbonate can be set up (Fig. 2).
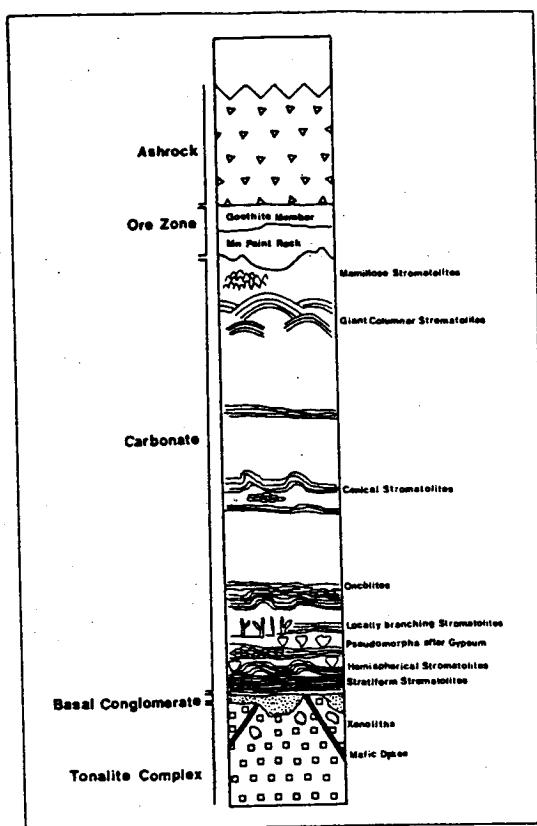
Figure 2:   Schematic stratigraphic column of the Steep Rock Group up to the
Ashrock Formation.   Note that thicknesses of individual members vary
greatly along strike.   Textural symbols not to scale.   From Wilks and
Nisbet, 1985 .

   Small scale stromatolites occur throughout the unit, but are best
developed near the base.   Here simple Stratifera-like stratiform structures
having flat to undulatory laminae develop into pseudocolumnar laterally-
linked structures.   These Irregularia-like structures pass upwards into
hemispherical laterally linked stromatolites.   Laminae are wavy .5-3.5mm,
and the structures are 5-15cm high and in basal diameter.   In places
branching walled and unwalled columnar forms occur, with heights up to
20cm.
   In the upper part of the Carbonate giant domal stromatolites occur.
These range from domed structures typically about 3m in diameter to tabular
bodies up to 5m or more long and .75m in stratigraphic height.   Near the
top of the unit, small mamillose stromatolites form an egg box fabric with
diameters up to 4cm and heights of 1.5cm.

Overlying the Carbonate is the Ore Zone which has been divided into a lower Mn Paint Rock Member and an upper Goethite Member (Jolliffe, 1955). The Mn Paint Rock (3%-18% Mn) is an earthy material with poorly developed varicoloured banding, made up of lumps of goethite, hematite, quartz and chert in a groundmass of the same minerals with calcite, kaolinite, pyrolusite and gibbsite. The contact with the underlying Carbonate is extremely irregular with pinnacles of carbonate protruding into the Paint Rock. This contact has been interpreted as an ancient karst surface (Jolliffe, 1955). The Mn Paint Rock passes sharply upwards into the Goethite Member (Mn < .3%) which is a predominantly brecciated lump ore of goethite (67%) and hematite (21%) with quartz and kaolinite.

Within the Ore Zone thin layers of Buckshot Ore occur. These layers comprise haematitic pisolites and fragments of haematite set in a lighter aluminous matrix of kaolinite and gibbsite. This material resembles a ferruginous bauxite in both outward appearance and chemical and mineral composition.

Overlying the Ore Zone is the Ashrock. The name refers to a high-Mg pyroclastic rock (22% MgO) which makes up to 90% of the unit. Interbedded within this are thin komatiitic basalt (15% MgO) lava flows.

Within the Goethite Member and Ashrock, pyrite lenses occur. These form discontinuous elongate bodies of massive pyrite closely associated with cherty and carbonaceous beds.

In contact with the Ashrock is the Metavolcanic Formation comprised of mafic and intermediate metavolcanics and clastic metasediments. This contact is nowhere exposed, and the Metavolcanics are thought to be separated from the underlying Ashrock by a structural break. At the present time they are provisionally included in the Steep Rock Group. The Group is interpreted as a sequence deposited in an extensional environment (a rift). With later extension and deformation the Marmion Complex and overlying rocks up to the Ashrock were tilted steeply to the west-southwest. The Metavolcanics, which are interpreted as deposits extruded in the centre of the rift were then folded and thrust up against the tilted succession. Regional lower greenschist metamorphism of the Steep Rock Group succeeded this deformation.

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